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The Clinton Street-Ballpark Aquifer in Binghamton and Johnson City, New York

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THE CLINTON STREET-BALLPARK AQUIFER IN BINGHAMTON AND JOHNSON CITY, NEW YORK

By

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U.S. Geological Survey

Prepared by
UNITED STATES DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

in cooperation with

NEW YORK STATE DEPARTMENT OF ENVIRONMENTAL CONSERVATION

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CONVERSION FACTORS

The following factors may be used to convert the conventional (English) units of measurement in this report to the International System of Units (metric system).

Multiply	by	To obtain
inches	25.4	millimeters
feet	.3048	meters
square feet	.0929	square meters
cubic feet	28.32 0.02832	liters cubic meters
miles	1.609	kilometers
square miles	2.590	square kilometers
feet per day	.3048	meters per day
feet squared per day	.0929	meters squared per day
cubic feet per second	28.32 0.02832	liters per second cubic meters per second
gallons	3.785	liters
gallons per minute	.06309	liters per second
gallons per day per square foot	40.7	liters per day per square meter
million gallons per day	3.785	million liters per day
degrees Fahrenheit (°F)	5/9(°F-32)	degrees Celsius (°C)

THE CLINTON STREET-BALLPARK AQUIFER

IN BINGHAMTON AND JOHNSON CITY, NEW YORK

by

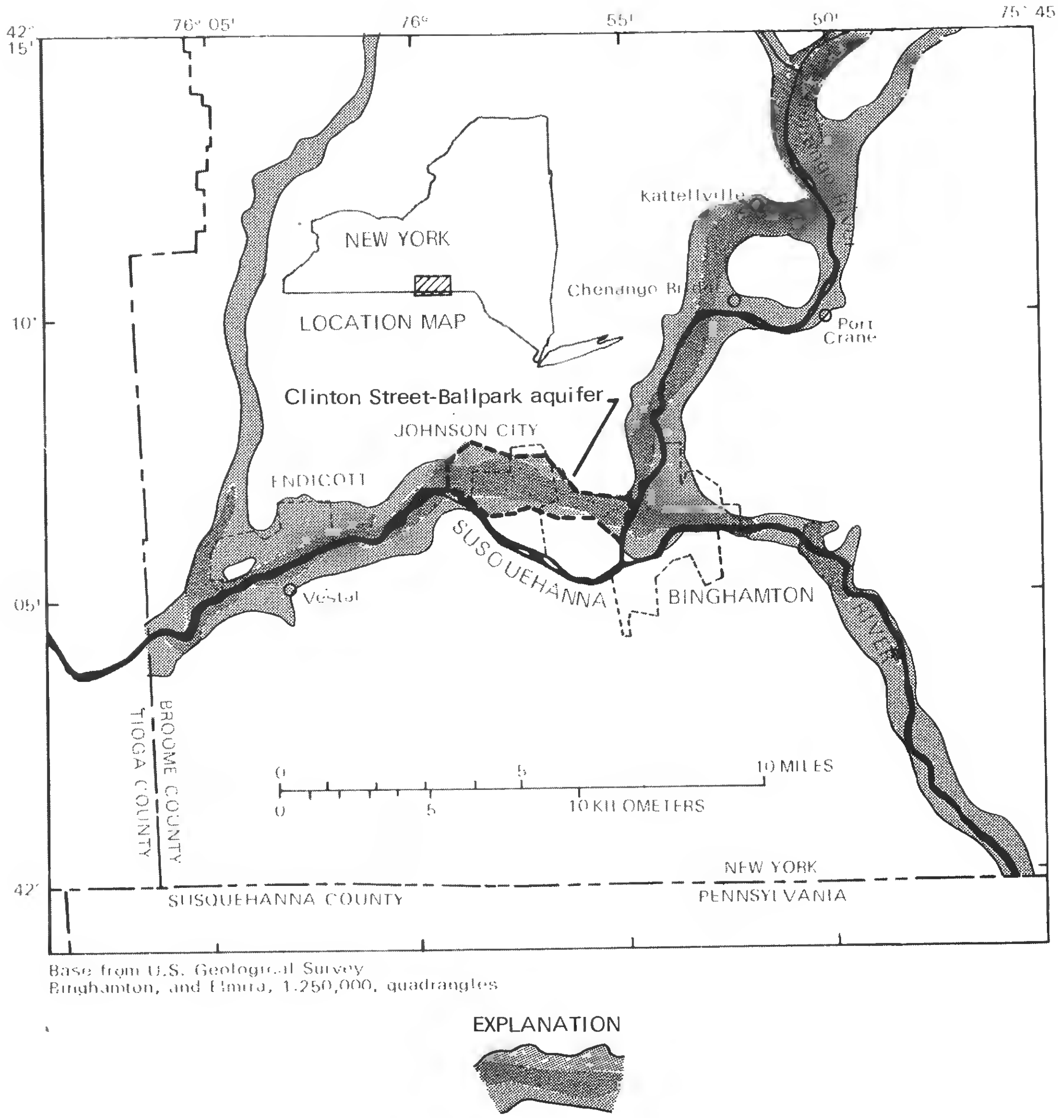
Allan D. Randall

ABSTRACT

The Clinton Street-Ballpark aquifer, which underlies 3 square miles (8 square kilometers) of urban land in the Susquehanna River valley, consists chiefly of permeable sand and gravel. Transmissivity generally exceeds 10,000 feet squared per day (900 meters squared per day). Lenses of silt locally restrict water movement. The east and west ends of the aquifer are in contact with two nearby rivers, but the central part of the aquifer is separated from the rivers by a ridge of till and bedrock. A combination of dry weather and relocation of pumping centers caused water levels to decline as much as 23 feet (7 meters) during the 1960's. These factors and increased use of chemicals for snow removal since the early 1950's are thought to be responsible for increases in hardness and chloride concentrations in water from heavily pumped wells.

The principal sources of recharge are (a) induced recharge from the Susquehanna and Chenango Rivers at the west and east ends of the aquifer--estimated annual potential is at least 7.6 billion gallons (29 billion liters); (b) precipitation on the aquifer and on thin sand and gravel bordering it--estimated average annual rate is at least 22 inches or 1.5 billion gallons (5.7 billion liters); and (c) infiltration from small streams crossing the aquifer--estimated average annual rate is 0.24 billion gallons (0.9 billion liters).

Urban land use and ground-water development together seem to have reduced evapotranspiration more than they have reduced recharge. Annual pumpage in 1967 was about 4 billion gallons (15 billion liters). Further ground-water development would require construction of additional wells near the ends of the aquifer to induce additional recharge from the rivers, and management of storage in the central part of the aquifer to supplement the river recharge. Additional wells in the central part of the aquifer may be desirable but will not increase the total amount of water available.



Aquifers composed of sand and gravel. Darker tone, more than 40 feet of saturated sand and gravel, locally overlapped by silt and clay. Lighter tone, generally less than 40 feet of saturated sand and gravel, overlying a thick section of silt and clay or overlying bedrock; thin sand and gravel aquifers locally beneath the silt and clay

Figure 1.--Location and geohydrologic setting of Clinton Street-Ballpark aquifer.

INTRODUCTION

This report describes and evaluates an important aquifer that extends from the western part of Binghamton through the central part of Johnson City, New York. An aquifer is a geologic formation that contains enough permeable material (in this case sand and gravel) to yield significant amounts of water to wells.

The report is divided into three major parts. The first and longest part was written to inform those in government and business who are responsible for planning and guiding future water-supply development near Binghamton. The reader who desires an understanding of the history and structure of the aquifer and of the factors that control well performance, replenishment of ground water, and water quality is referred to the entire first (nontechnical) part, whereas the reader who wishes to consider only the future potential of the aquifer may turn to the section "Conclusions Related to Aquifer Management" and to table 2, which summarizes recharge rates.

The second part consists of technical appendices that describe the records, assumptions, methods of computation, and problems involved in analyzing and interpreting data for this report. The appendices are intended for use by hydrologists who wish to evaluate the estimates of aquifer yield presented and to prepare specific recommendations consistent with their own knowledge and objectives.

The final part is a series of maps of the aquifer. They are referred to throughout the report but are placed together at the end as plates 1-8 for easy comparison.

LOCATION AND DEFINITION OF THE AQUIFER

The most productive aquifers in the part of New York drained by the Susquehanna River are composed of sand and gravel, are interbedded with silt and clay in varying proportions, and border on major streams. In many places, water can easily move back and forth between the aquifers and the streams. However, in the western part of the city of Binghamton and in Johnson City, where the Susquehanna River flows very close to the south side of its valley, low hills of till or bedrock form an impermeable wall 3 miles (5 kilometers) long that separates the river from a sand-and-gravel aquifer (fig. 1).

This separated aquifer is referred to in this report as the Clinton Street-Ballpark aquifer. Clinton Street is a major avenue in Binghamton that approximately follows the aquifer. The ballpark in Johnson City, which was once the home of the Binghamton Triplets baseball club, was destroyed in 1969 to make room for a new highway, but the adjacent "Ballpark well" remains a major source of municipal water supply.

The Clinton Street-Ballpark aquifer is bounded on the north by the impermeable bedrock wall of the Susquehanna valley, on the south by low hills of impermeable till and bedrock, on the east by the Chenango River, and on the southwest by the Susquehanna River. On the northwest, no natural boundary was recognized, so an arbitrary boundary was drawn north from the northernmost river bend at Johnson City. Aquifer materials continue east and west beyond the rivers (fig. 1). The lower boundary of the aquifer is defined in plate 2 by contours on the underlying bedrock and till surfaces.

HISTORY OF DEVELOPMENT AND STUDY OF THE AQUIFER

Ground water has been used extensively in the Triple Cities (Binghamton, Johnson City, and Endicott) since 1912, when the village of Endicott began to pump a cluster of wells by compressed air. Withdrawal from the Clinton Street-Ballpark aquifer began during 1928-31 with construction of the Ballpark well and four other wells to supply Endicott-Johnson Corporation and the village of Johnson City. During the mid-1930's, Ansco (now GAF) Corporation established a well field north of Clinton Street in Binghamton, as did International Business Machines Corporation in Endicott. All these well fields have since been enlarged.

The U.S. Geological Survey made brief reconnaissances of ground water in the Triple Cities in 1942 and 1945 and prepared a report that describes each well field, analyzes trends in pumpage and water levels, and presents logs and specifications of wells (Brown and Ferris, 1946). The authors recognized the barrier between the Clinton Street-Ballpark aquifer and the Susquehanna River and noted that increasingly large pumpage had caused a persistent decline in ground-water levels (Brown and Ferris, 1946, p. 23-25). To monitor future water-level trends, the Geological Survey drilled 23 observation wells in 1946 in Binghamton and Johnson City, most of them within the Clinton Street-Ballpark aquifer. Continuous records of water levels in three wells were obtained for a year or more and were published (U.S. Geological Survey, 1951-67). Water levels were measured intermittently in the remaining wells, and these records are available from the Geological Survey office in Albany, N.Y.

In 1965, the New York State Conservation Department (now Department of Environmental Conservation) and the U.S. Geological Survey began a cooperative study of the Susquehanna River basin. Included in this study was an evaluation of the Clinton Street-Ballpark aquifer. This aquifer received special attention because (1) it was the source of all water used by Johnson City and much of the water used by the largest industry in Binghamton, and (2) the combination of numerous observation wells, many water-level measurements previously obtained, separation of the aquifer from the river, and heavy pumpage of ground water together provided an unusually favorable situation in which to evaluate ground-water recharge in an urban area. The results may be applied in aquifer evaluations elsewhere in the Susquehanna basin. Most of the information collected from 1965 through 1968 was obtained through the cooperation of Edward Tomic, Bert Davis, and others at the Johnson City Water Department, and Vincent Storman, Hans Wagner, and others at GAF Corporation.

GEOLOGIC FRAMEWORK

The Clinton Street-Ballpark aquifer was formed about 17,000 years ago (Cadwell, 1973) as the last glacier retreated from south-central New York. Deep valleys, originally carved by streams, had been widened and deepened by tongues of ice (Coates, 1966). While the glacier was melting, lakes continually formed between the remaining ice and older sediment downvalley. Turbulent rivers of meltwater built deltas of sand and gravel where they entered these lakes, and the silt, clay, and very fine sand they carried in suspension settled to the lake bottoms. Much of the sediment that was deposited west of the Chenango River in Binghamton and in Johnson City is permeable sand and gravel that today forms the Clinton Street-Ballpark aquifer.

The successive geologic units that compose and border the aquifer are described in table 1; the diagram in figure 2 illustrates the arrangement of these units.

Glacial deposits in the Susquehanna River basin range from "bright" to "drab" (Denny and Lyford, 1963; Moss and Ritter, 1962). The bright deposits contain fragments of many different rock types from remote locations and thus have a colorful appearance; the drab deposits are derived almost entirely from local shale bedrock. Near Binghamton, the drab glacial sand or gravel deposits are slightly older than the bright ones; that is, the bright overlies the drab wherever both types are present (Randall, in press). The change is commonly gradational over many feet if no fine-grained beds intervene. Because small tributary streams continued to bring drab gravel into the major valleys after the retreat of the ice, thin postglacial drab gravel may overlie bright glacial gravel near such streams. Geologists may find these relationships useful in tracing units from one borehole to another.

Distribution of the various geologic units at land surface is shown in a surficial geologic map (plate 6). Their structure and position below land surface are illustrated in figure 3.

ABILITY OF THE AQUIFER TO TRANSMIT WATER TO WELLS

The concept of transmissivity is used by hydrologists to express in quantitative terms the ability of aquifers to transmit water. Transmissivity is a measure of the rate at which water would flow through a vertical strip of specified width extending from the top to the bottom of the aquifer, assuming a 1/1 hydraulic gradient. A 1/1 gradient, which means a 1-foot decline in water level for each foot of water movement, is steeper than gradients usually observed in aquifers but serves as a standard for comparison. However, even though transmissivity is defined exactly and expressed numerically, it is difficult to measure precisely in most glacial aquifers because it varies widely from place to place.

The Clinton Street-Ballpark aquifer is composed mostly of permeable materials. Transmissivity in the central part of the aquifer generally exceeds 10,000 feet squared per day and locally may reach 100,000 feet squared per day (900 to 9,000 meters squared per day).

Table 1.--Geologic units in and near the Clinton Street-Ballpark aquifer

Geologic unit (youngest to oldest)	Number in figures 2 and 3	Lithology (materials composing unit)	Distribution, thickness, and position	Hydrologic significance
Fill	8	Chiefly trash and ashes; some sand, gravel, and other materials	Most natural depressions in Binghamton and Johnson City have been raised 5 to 20 feet by fill; some are now unrecognizable.	Not tapped by wells. Increases dissolved-solids concentration and acidity of infiltrating water, but effect decreases as age of fill increases.
Flood-plain silt	7	Brown silt and very fine sand with roots and a little fine organic matter.	Mantles lowlands inundated during major floods; typically 5 to 15 feet thick. May rest on all older units (1-5).	Not tapped by wells. Poorly permeable; limits recharge of underlying aquifers from floodwater and possibly from heavy rainfall.
Alluvial fan deposits	6	Gravel, moderately sandy and in general moderately silty. Most stones are flat pieces of local shale or siltstone.	Deposited by small streams where they enter the Susquehanna valley. May rest on all older units (1-5).	Permeable, but too thin to supply large-capacity wells. Water from small streams infiltrates through alluvial fan deposits to stratified glacial deposits.
Older river alluvium	5	Sand and gravel, bright but leached partially to completely free of limestone.	Interfingers with and overlies late-glacial lakebeds near Chenango River; as much as 35 feet thick. Relation to other units uncertain. May cap stratified glacial deposits beneath flood-plain silt elsewhere, but is not recognized or mapped.	Highly permeable and in good hydraulic contact with Chenango River. Could be tapped by large-capacity wells.
Late-glacial lakebeds	4	Silt and very fine sand with some clay and scattered tiny plant fragments; commonly grades into peat or highly organic silt at top.	Fills irregular depressions left when ice blocks melted, chiefly in a narrow east-west zone near deepest part of bedrock valley; as much as 80 feet thick. Generally overlies bright gravel.	A significant barrier to infiltration and ground-water flow in many places.
Stratified glacial deposits	3			
Bright gravel	3c	Sandy gravel and pebbly sand containing variable amounts of silt; highly calcareous. Upper part very bright (35 to 75 percent of the pebbles are limestone and other rock types not derived from local bedrock). Lower part moderately bright (15 to 30 percent exotic pebbles).	Present over much of the valley as broad terraces or underlying younger units; thickness varies widely, locally exceeds 100 feet.	Highly permeable, tapped by several large-capacity wells, but locally above water table. The abundant limestone in this unit causes water that migrates through it to have high hardness (250-400 milligrams per liter).
Lake beds	3b	Silt to fine sand, some clay, no plant fragments.	Lenses may occur anywhere within unit 3, but seem to be most common between the bright and drab gravels.	A significant barrier to infiltration and ground-water flow in places.

Drab gravel	3a	Sandy gravel and pebbly sand with variable amounts of silt; weakly calcareous. Pebbles are almost entirely local shale and siltstone, with 10 percent or less exotic rock types.	Present at land surface along north and south sides of valley; commonly underlies bright gravel (directly or with intervening lake beds) in central part of valley; varies widely in thickness.	Highly permeable; tapped by several large-capacity wells.
Glacial till	2	Mixture of silt, clay, gravel, and sand, tough and compact; commonly called hardpan. May contain minor sand and gravel lenses.	Immediately overlies bedrock. Only about 1 foot thick in places, but forms low hills in southern part of Susquehanna valley.	Very poorly permeable. Low hills of till prevent movement of water between aquifer and Susquehanna River for 3 miles west from Chenango River
Bedrock	1	Interbedded shale and siltstone.	Present everywhere beneath other units.	Poorly permeable; serves as north, and part of south, aquifer boundary, but yields 100 to 300 gallons per minute of salty water to wells several hundred feet deep.

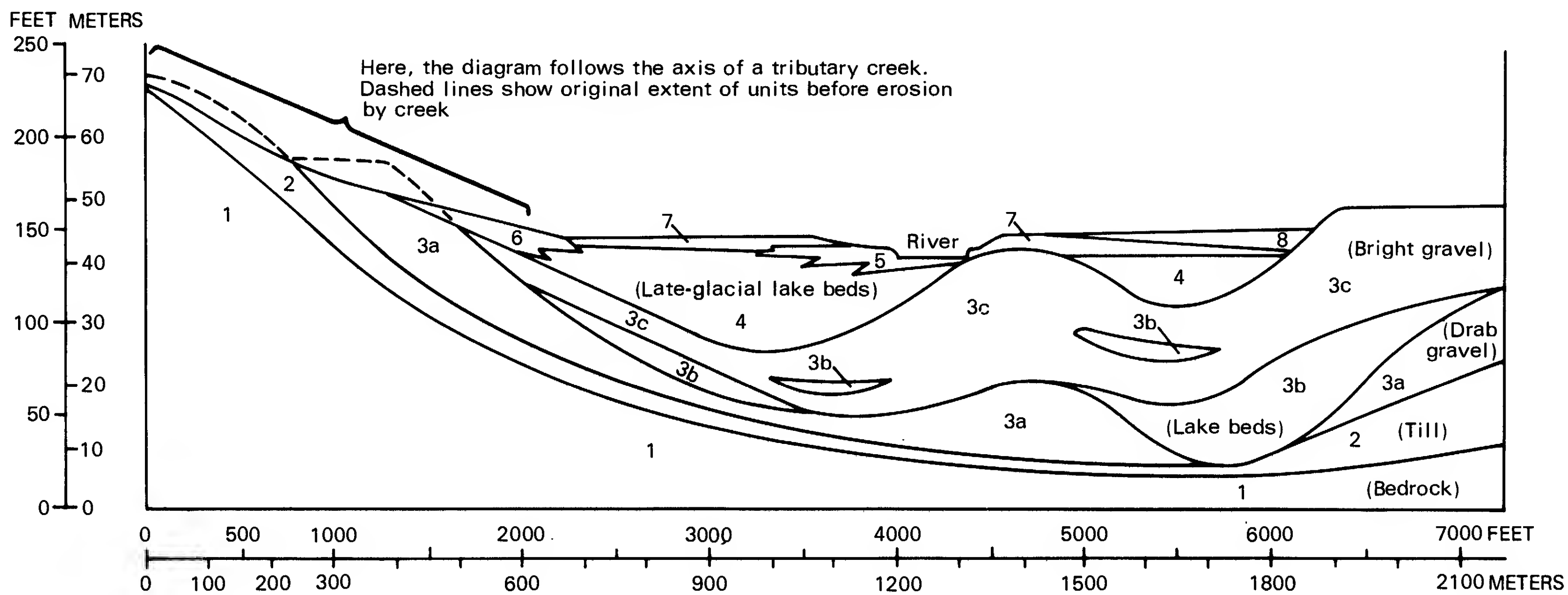


Figure 2.--Idealized diagram illustrating arrangement of geologic units numbered and described in table 1. Note that diagram would appear much flatter if drawn to same scale vertically and horizontally.

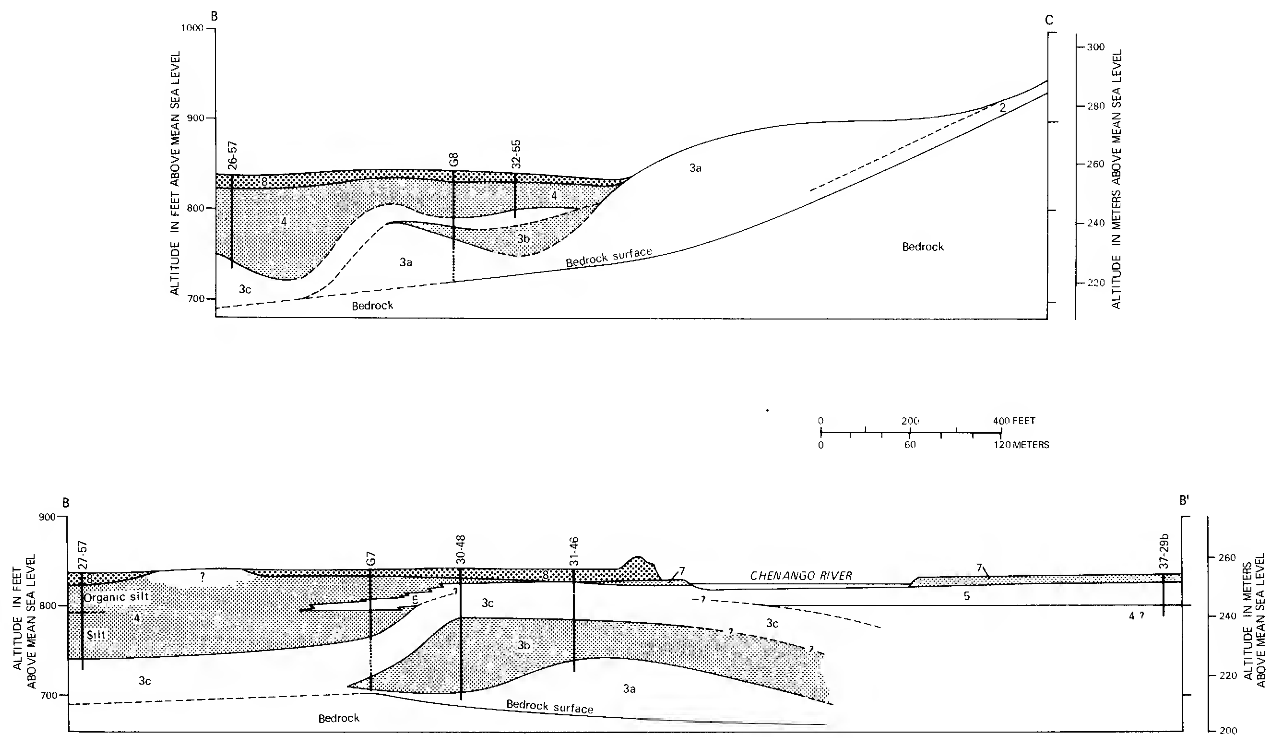


Figure 3.--Geologic sections within the Clinton Street-Ballpark aquifer. See plate 6 for location of sections in relation to surficial geology and plate 1 for location of some individual wells.

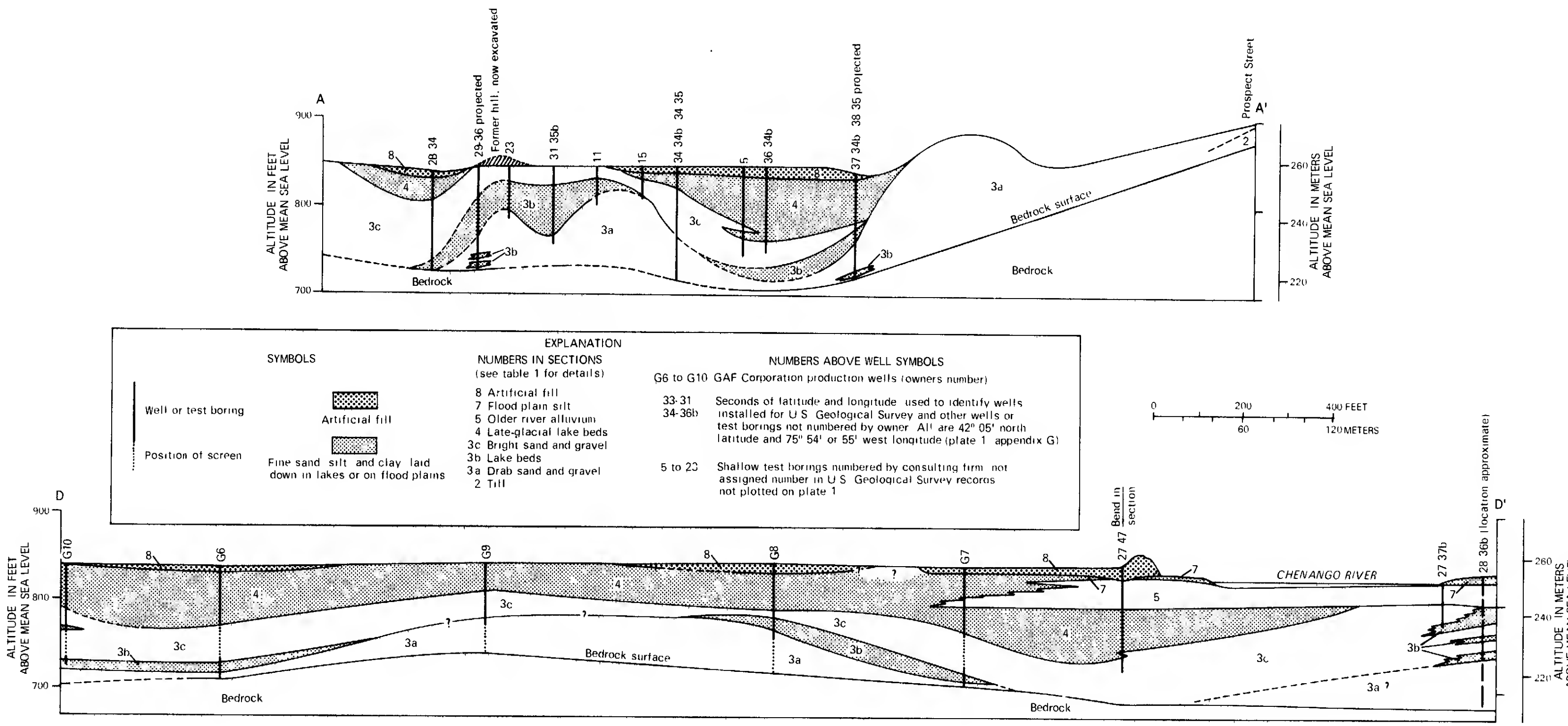


Figure 3 (continued).--Geologic sections within the Clinton Street-Ballpark aquifer.

A map of transmissivity (plate 3) shows the major trends as indicated by approximate values computed for 100 unevenly distributed points, as explained in appendix A. Despite its limitations, a generalized map such as plate 3 is useful and can be refined as data become available from new wells or from tests of aquifer models. For example, knowledge of the variation in transmissivity from place to place can be helpful in selecting sites for future wells, inasmuch as well yield for short periods of pumping is approximately proportional to transmissivity. That knowledge, together with other data, can also provide a starting point for predicting how large a decline in water level should result from pumping one or more wells for specified periods of time. Hydrologists can make such predictions by using various well-known equations (Ferris and others, 1962; Walton, 1962) with a simplified diagram of aquifer dimensions, or they can make more precise predictions by using complex electric-analog or digital-computer models (Pinder and Bredehoeft, 1968).

The yield of an aquifer as a whole depends not only on how rapidly it can transmit water to each individual well, but also on how much water is stored in it and how much enters it naturally and by artificial means. These factors are discussed in the next two sections, "Water levels and storage in the aquifer" and "Recharge of ground water."

WATER LEVELS AND STORAGE IN THE AQUIFER

Historical Account

Before large-scale ground-water development began, water in the Clinton Street-Ballpark aquifer flowed toward, and eventually seeped into, the Susquehanna and Chenango Rivers. The exact shape of the natural water table is not known, but it must have sloped gently east and west from a divide near the city line at an altitude of about 840 feet (256 meters). From 1946 (the first year in which water levels were measured at numerous locations) through 1958, most of the water in the aquifer flowed toward pumping centers at Charles Street in Binghamton and Camden Street in Johnson City (plate 7).

Two factors caused water levels to change significantly in the 1960's. In 1959, Johnson City began large-scale use of wells 4-6 and the Ballpark well, all in the central part of the aquifer, and at the same time reduced pumpage from wells 1-3 at Camden Street near the Susquehanna River (fig. 4). Also, from 1962 through 1967, an unusually persistent and occasionally severe drought affected northeastern United States (Barksdale and others, 1966). Rainfall near Binghamton was from 10 to 20 percent below the long-term average for each of these 6 years, and the percentage deficit in runoff was even greater (fig. 5). The amount of rainfall that infiltrated into the Clinton Street-Ballpark aquifer during these years must also have been much below normal, but pumpage remained constant until 1965 (fig. 4). Because of the reduced ratio of recharge to withdrawal and the relocation of pumping centers, water levels declined substantially in the central part of the aquifer. Comparison of water-level contours for September 25, 1958 (plate 7) with those

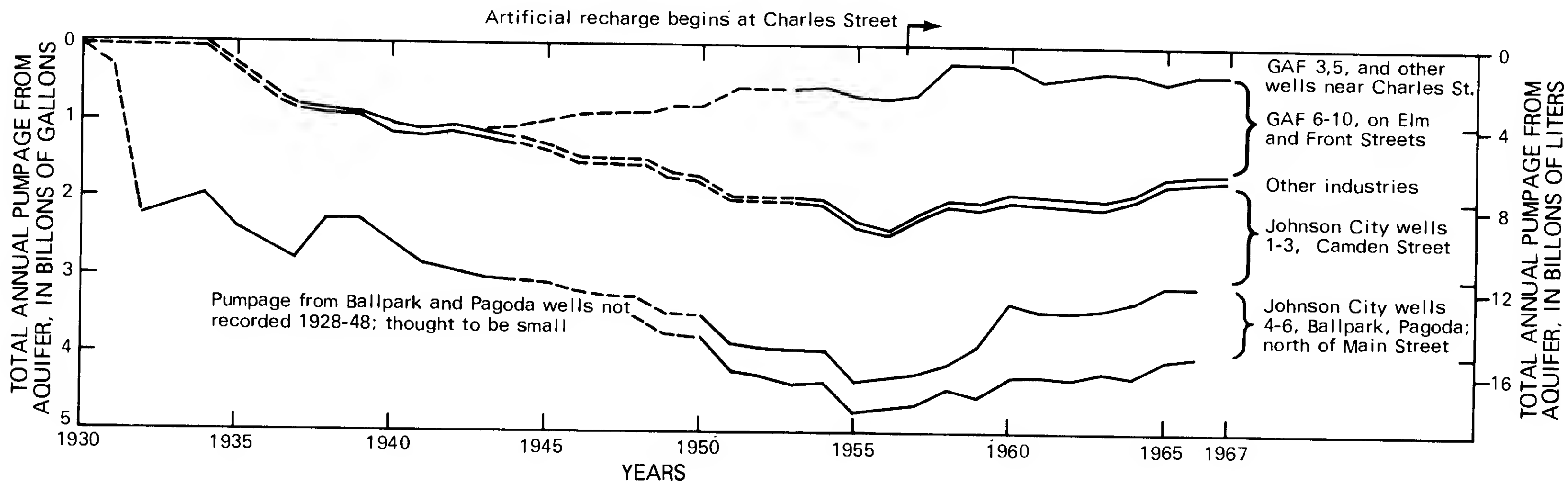


Figure 4.--Pumpage from the Clinton Street-Ballpark aquifer, 1930-67. Solid lines indicate pumpage according to records of well owners; dashed lines are estimates. Pumpages shown for GAF wells near Charles Street are actual pumpage less any artificial recharge. More complete data for 1958-67 are given in appendix B; for 1937-44, in Brown and Ferris (1946).

for October 6, 1967 (plate 8), shows that water levels dropped approximately 10 feet (3 meters) in Binghamton and as much as 23 feet (7 meters) in Johnson City, except near the rivers at either end of the aquifer. In other words, the amount of water stored in the aquifer was reduced by about 1 billion gallons (4 billion liters) between 1958 and 1967. Water-level records suggest that most of the decline took place between 1962 and 1965 (fig. 6). The decline would have been greater had some users not reduced water use in 1965 because they were concerned about the low water levels in production wells.

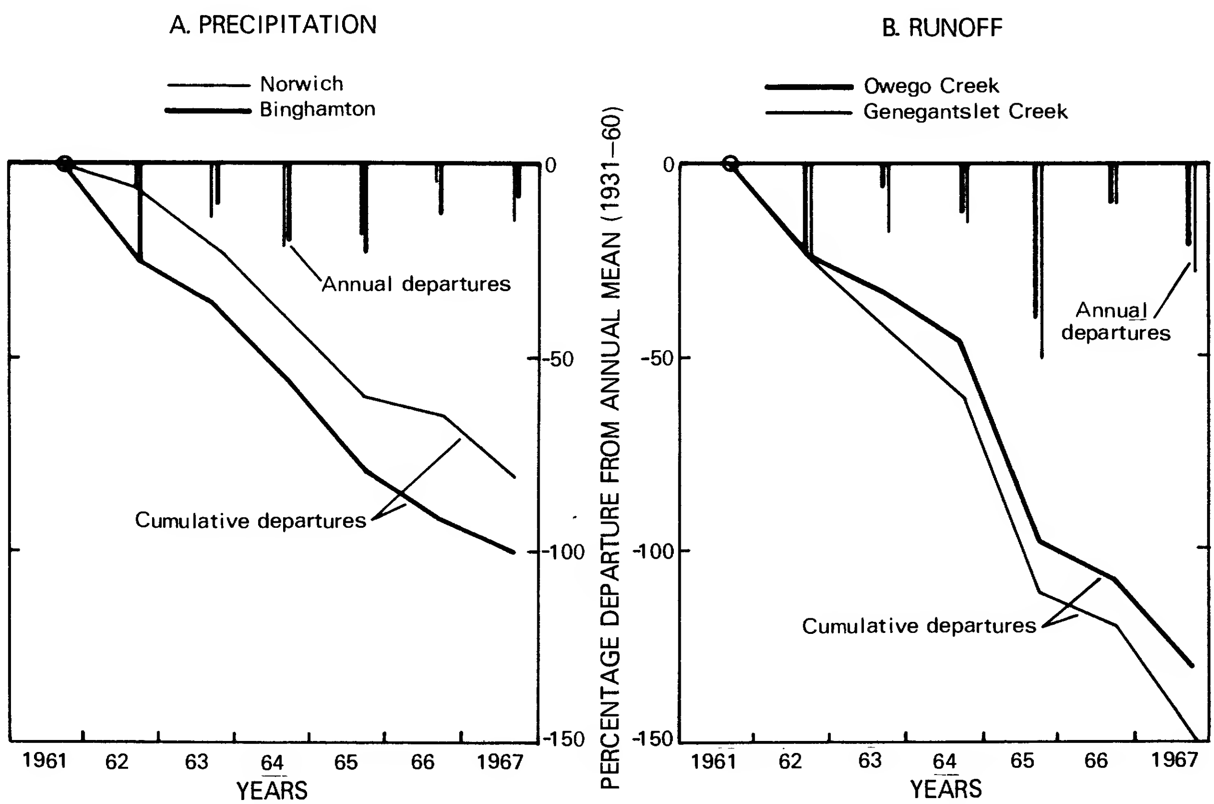


Figure 5.--Departures from normal rainfall and runoff during the 1962-67 drought.

A) Precipitation at Binghamton, near the Clinton Street-Ballpark aquifer, and at Norwich, 35 miles (56 kilometers) northeast of Binghamton.

B) Runoff from Oswego Creek, 18 miles (29 kilometers) west of Binghamton, and from Genegantslet Creek, 22 miles (35 kilometers) north of Binghamton. The cumulative runoff departure from September 30, 1961 through September 30, 1967 was equal to 1.3 and 1.5 years of normal runoff in these two basins, and to 1.5 years runoff in the entire eastern Susquehanna River basin in New York (not shown).

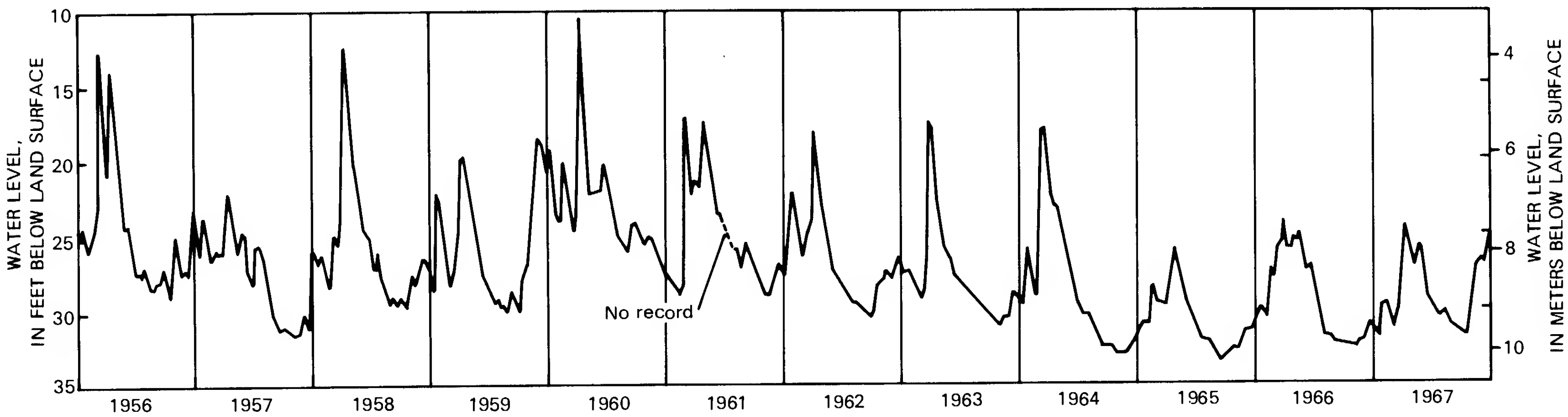


Figure 6.--Water levels in U.S. Geological Survey observation well in Johnson City, 1956-1967. This well is near the west end of the Clinton Street-Ballpark aquifer (lat 42°06'57 N.", long 75°58'35 W." in plate 1), where there is no ridge of till separating the aquifer from the Susquehanna River, and is approximately 1,000 feet (300 meters) north of Johnson City production wells 1, 2, and 3, whose average annual production rate decreased 25 percent after 1959. Therefore, although water levels during the drought of the 1960's were the lowest since records began in 1947, the water-level decline here was not as great as farther east in the aquifer.

Water Use

For many years, half the water withdrawn from the Clinton Street-Ballpark aquifer has been used for public supply by Johnson City and its satellite water districts. The other half has been used by GAF Corporation and other industrial firms, principally for industrial cooling but also for washing film, for irrigation, and for air conditioning.

Annual pumpage since 1930 is given in figure 4; pumpage from 1958 through 1967 is tabulated for each individual well or well field in appendix B. Daily withdrawals by all users are greatest during the summer.

Storage Available

The Clinton Street-Ballpark aquifer may be used as a reservoir, which, although underground, is usable in much the same way as a surface reservoir. From the array of wells and the pattern of withdrawals in 1967, the usable reservoir storage capacity is inferred to be 1,320 million gallons (5 billion liters). If management were to establish a more even distribution of wells and allow most wells to remain idle for several months each year, usable storage capacity would be at least 1,700 million gallons (6.4 billion liters). The basis for these storage estimates is given in appendix C.

RECHARGE OF GROUND WATER

Any aquifer evaluation must consider how water enters (recharges) the aquifer, how it leaves the aquifer, and how much of this water can be captured by pumping. Under natural conditions, nearly all water in the Clinton Street-Ballpark aquifer originated as precipitation on the land surface directly above the aquifer. After percolating downward to the water table, the water generally flowed to the east or west until it reached and seeped into the Chenango or Susquehanna River. Since at least the late 1940's, however, the water table has been lowered below river level in enough places that ground water no longer seeps into the rivers but, instead, leaves the aquifer almost entirely through pumped wells. A minor amount is used by the relatively few plants whose roots reach the water table. Much of the ground water still originates from local precipitation, but now a larger amount infiltrates from streams.

The pattern of recharge and discharge in 1967 is shown in figure 7. The sources of recharge are described in the following paragraphs, and the amount of water available from each source is listed in table 2. Appendices D, E, and F explain how each source was evaluated.

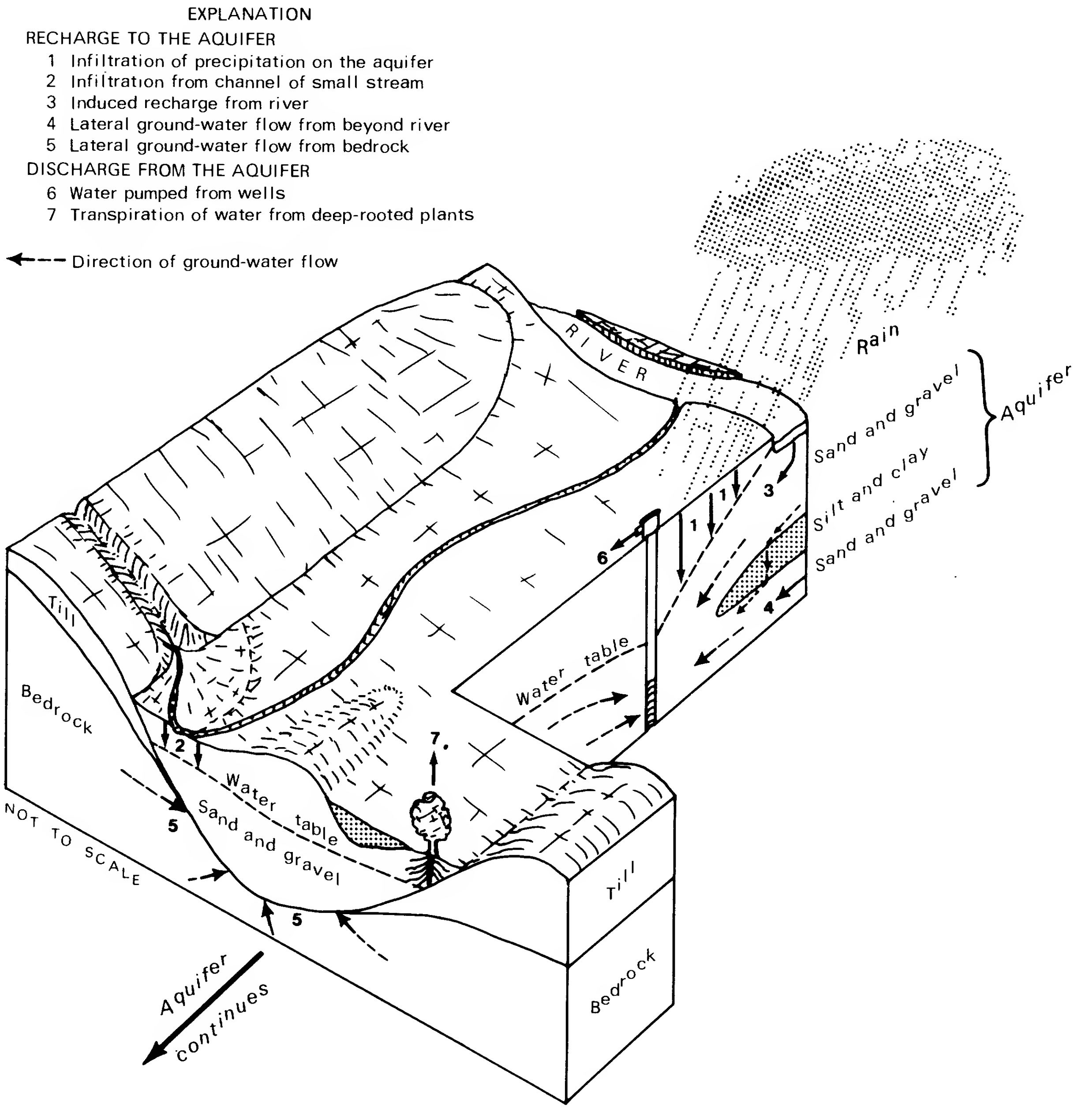


Figure 7.--Idealized diagram of recharge to and discharge from the Clinton Street-Ballpark aquifer as of 1967.

Recharge from rainfall and snowmelt

The Clinton Street-Ballpark aquifer underlies an urban area that is occupied primarily by free-standing houses but also includes some large commercial and industrial buildings. In the 1960's, only two large tracts of open land remained in the area; together they covered approximately 14

Table 2.--Recharge to the Clinton Street-Ballpark aquifer

[in millions of gallons]

	During period of study		Estimated potential	
	October 1966-67 (adjusted to 1 year)	May 1967 to October 1967 (164 days)	Long-term median year	Lowest year of typical 30-year period
Total recharge to aquifer	4,200	1,500	9,300	8,800
From Chenango and Susquehanna Rivers	2,500	1,000	1/7,600	1/7,600
From Little Choconut Creek	225	115	235	160
From precipitation 2/	1,440	380	1,510	1,030

1/ These rates could be achieved only by placing additional wells near the Susquehanna and Chenango Rivers, as described in the section "Conclusions related to Aquifer Management" and in appendix F. Minimum river flow exceeds potential infiltration; thus, estimates for a dry year are the same as for a median year. Reduction in flow of the Chenango River would be less than 25 percent of the lowest 7-day average flow for a typical 30-year period (Ku and others, 1975); reduction in flow of the Susquehanna River would be less than 2 percent under comparable low-flow conditions.

2/ Calculations explained in appendix E led to two estimates of recharge from precipitation--one based on wells in Binghamton, another on wells in Binghamton and Johnson City. Values in table 2 and in most computations in this report use the lower estimate to provide conservative predictions. If the higher estimate were correct, recharge from precipitation would be raised to 1,750, 460, 1,850, and 1,300 million gallons; corresponding decreases would be required in river recharge (first two columns) and corresponding increases in total recharge (last two columns).

percent of the aquifer. Possibly 20 to 30 percent of the land surface above the aquifer was covered by streets, paved parking lots, and buildings. Even so, most precipitation infiltrated into the aquifer, including some that was discharged onto the land surface from roof gutters or diverted to open-bottomed catch basins.

Recharge from precipitation can be calculated only after accounting for (1) recharge from small streams, rivers, and other minor sources, and (2) net change in the amount of water stored in the aquifer during the period selected for study. Discharge from the aquifer is easily determined because pumpage records are available for most wells. By subtracting from known discharge any recharge from streams or other sources and any increase in storage, or by studying localities where these factors have no effect on known discharge, one can calculate the quantity of recharge from precipitation. In an average year, recharge from precipitation is at least 22 inches (560 millimeters) and may be as much as 27 inches (690 millimeters). It is probably as little as 15 inches (380 millimeters) once in 30 years.

Recharge from small streams

During the 1960's, the water table in the Clinton Street-Ballpark aquifer was several feet below the channels of small streams such as Little Choconut Creek and Trout Brook. Consequently, some water from the streams must have infiltrated to the underlying aquifer. Glenwood Creek and Trout Brook are encased in paved channels or culverts where they cross the aquifer, so loss of water by infiltration from these streams was probably very small and would be difficult to study; therefore, it has been ignored in this report. Loss of water from Little Choconut Creek and Finch Hollow Creek in Johnson City was easily observed because some reaches of both streams were occasionally dry in late summer, and other reaches would have been dry were it not for water discharged from industrial sewers. To evaluate this source of recharge, the flow of Little Choconut Creek and its tributaries over and near the aquifer was measured on 16 occasions from 1966 through 1968. Although several difficulties limited the accuracy of the measurements, the long-term average rate of recharge is probably close to 235 million gallons (890 million liters) per year.

Induced recharge from rivers

Heavy pumping near both ends of the Clinton Street-Ballpark aquifer has lowered water levels several feet below river level and thereby has induced river water to infiltrate into the aquifer. Most of this induced recharge is captured by wells near the rivers. Wells that capture induced recharge can be distinguished from those that do not by means of flow nets (plates 7 and 8) because the flow lines show the directions of ground-water movement. The amount of induced recharge from rivers can be estimated by first determining the rate of recharge from precipitation in the part of the aquifer that is not affected by induced recharge, then subtracting this rate, along with recharge from any small streams, from the total pumpage in the remainder of the aquifer. A less precise method of estimating induced recharge is to compare temperature or chemical quality of water pumped from wells with that of river water and to calculate what proportions of river water and native ground water would produce the observed temperature or quality at each well.

Recharge from lateral ground-water flow and other sources

Some ground water doubtless flows into the Clinton Street-Ballpark aquifer laterally from areas of sand and gravel beyond the Chenango River in Binghamton and from similar areas near the Susquehanna River in Johnson City west of the aquifer (fig. 1). Temperature fluctuations in water flowing westward about 100 feet (30 meters) beneath the Chenango River show that part of that water is induced recharge, which must have infiltrated near the east bank of the river and presumably mixed with ground water from beyond the east bank. Ground-water flow into the aquifer from sand and gravel beyond or near the rivers is thought to be a minor percentage of recharge under 1967 conditions and would be replaced by additional recharge from the rivers if it ceased to be available; therefore, it is included in estimates of induced recharge.

A small amount of ground water enters the Clinton Street-Ballpark aquifer from the poorly permeable till and bedrock on the south and especially the north side, and some recharge may result from leaky water mains or sewers. All such recharge is included in estimates of recharge from precipitation.

WATER QUALITY

Temperature

Temperature of ground water in the central part of the Clinton Street-Ballpark aquifer is near 11° or 12°C (52° or 53°F) at depths tapped by most production wells. Variation in temperature from winter to summer is no more than 1°C (1° or 2°F) in most places. Vertical temperature profiles obtained in a few wells in the central part of the aquifer showed that the water was slightly cooler near the bottom of the aquifer than near the water table throughout the year. Perhaps the many heated buildings above the aquifer serve to keep the uppermost ground water relatively warm even in winter.

Temperature of water in the Chenango and Susquehanna Rivers fluctuates widely, from 0°C (32°F) in midwinter to 27°C (81°F) in late summer. This fluctuation is reflected in the large seasonal changes in temperature that occur underground near the ends of the Clinton Street-Ballpark aquifer, especially in highly permeable layers of gravel that are in direct contact with the riverbed and transmit large amounts of induced recharge. For example, water pumped from GAF well 7, 450 feet (135 meters) from the Chenango River, has fluctuated from 6°C (42°F) in early April to about 17°C (63°F) in early September for many years. Water in the most permeable layers penetrated by U.S. Geological Survey test well 27-47, 150 feet (45 meters) from the river, has fluctuated from 1°C (34°F) to 22°C (71°F)--nearly as wide a range as that of the river. Locations of these wells are shown in plate 1.

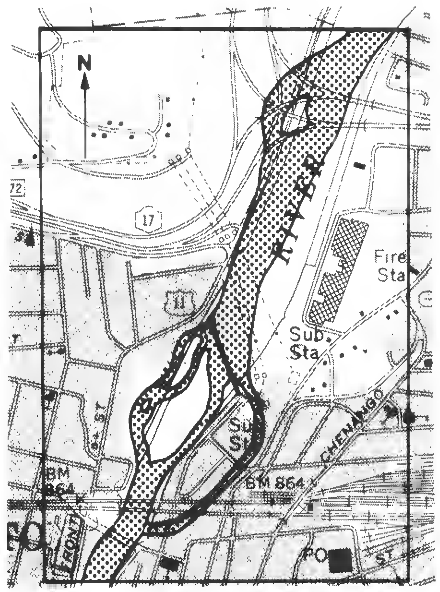
Sanitary Quality

The sanitary quality of water from Johnson City municipal wells has been monitored by the New York State Department of Health for many years. Regular analyses have never detected coliform bacteria. It is significant that Johnson City wells 1-3 produce bacteria-free water because about half the water reaching these wells infiltrates from the Susquehanna River, which until 1968 received Johnson City's sewage untreated only 100 feet (30 meters) upstream. The river also received Binghamton's sewage untreated at several points a few miles upstream until 1960, then from 1960 through 1968 received primary-treated effluent only 1.2 miles (1.9 kilometers) upstream. Presumably, fine-grained sediments in the riverbed and in the underlying aquifer filter bacteria out of the infiltrating river water. An incident in the late 1950's suggests that the thin, near-surface layer of fine-grained sediment in this locality (plate 6) limits infiltration and thus may also limit bacteria movement. A water-treatment plant operator at Johnson City recalled that one summer it was necessary to dig up a large water main a short distance east of wells 1-3 along the lower course of Little Choconut Creek. Water from the creek flowed directly into the excavation. Just upstream, the creek received a large discharge of hot water from the Goudey Station powerplant. Well 2 was in operation, and within several hours, temperature of water pumped from the well rose alarmingly, perhaps by as much as 11°C (20°F). Apparently the excavation cut through the shallow fine-grained sediment into gravel, through which heated water infiltrated rapidly to well 2. Well 2 was shut down, and well 3, 120 feet (37 meters) distant, was started, but no increase in temperature was noted at well 3 before the excavation was filled in a day or so later (Bert Davis, oral commun., 1966).

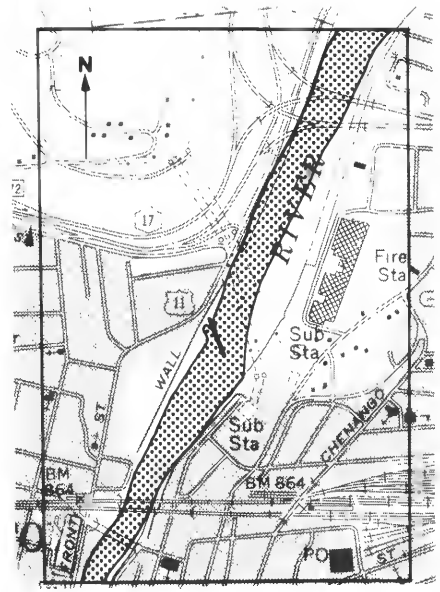
The vast majority of municipal and industrial wells tapping unconsolidated materials in New York, including wells fed by induced recharge from nearby rivers, produce bacteria-free water. However, coliform bacteria from the Susquehanna River have frequently reached a municipal well several miles west of Johnson City (Randall, 1970). The riverbed has been excavated repeatedly near that well, and the disruption or removal of natural layers of fine-grained sediment in the riverbed may have permitted water to enter the underlying aquifer more rapidly, resulting in migration of bacteria to the well. The east end of the Clinton Street-Ballpark aquifer may also be susceptible to migration of bacteria underground because the highly permeable aquifer is in immediate contact with the Chenango River where the riverbed has been repeatedly excavated and partially relocated, most recently in 1966 (fig. 8). The numbers of bacteria in river water at the east end of the aquifer in future years should remain low in comparison with those formerly observed west of Johnson City, because the east end of the aquifer is upstream from the heart of the Triple Cities and because of modern sewage treatment. Nevertheless, under conditions of maximum ground-water development, some coliform bacteria may be expected to penetrate a few hundred feet into the aquifer from the river.

Chemical Quality

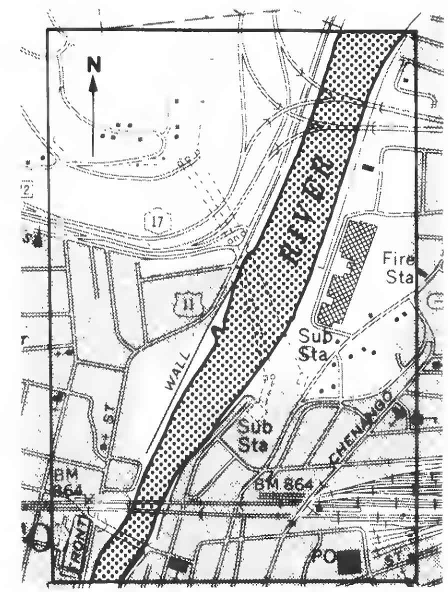
The most abundant dissolved chemical constituents in water in the Clinton Street-Ballpark aquifer are calcium and bicarbonate. The water is generally suitable for municipal and some industrial uses, although some



About 1912



1961



1968

For many years the Chenango River occupied multiple channels below the Cutler dam, which diverted water into the eastern channel around a feature then known as Noyes Island. The western channels were cut off in 1913 when the original Front Street dike was built. In 1938 the U.S. Army Corps of Engineers raised the dike, cut off the eastern channel, and excavated the main channel as part of a flood-control program.

In 1965 and 1966, the New York State Department of Transportation shifted the channel eastward north of the old Cutler dam to accomodate new highways, and deepened the channel as much as 4 feet for 900 feet downstream.

Figure 8.--Changes in location of Chenango River channel. (From U.S. Geological Survey topographic quadrangle maps, 1935, 1961, and 1968; and from old map in files of Binghamton City Engineer.)

samples have slightly exceeded the limits for iron, manganese, and dissolved solids suggested by the U.S. Public Health Service (1962) for drinking water. Nitrate generally ranges from 0.0 to 1.5 milligrams per liter as nitrogen, well below the suggested limit. Analyses of water from 30 wells in this aquifer, tabulated by Randall (1972), reveal two consistent patterns:

(1) Water is more mineralized and much harder in the central part of the aquifer than near the ends. Hardness is generally 300 to 400 milligrams per liter, but near the ends of the aquifer it is more commonly 150 to 210 milligrams per liter. Chloride is generally 30 to 60 milligrams per liter, but near the ends of the aquifer it is only 10 to 15 milligrams per liter (plate 4).

(2) Mineral content increased during the late 1950's and the 1960's. Increases in hardness and chloride are illustrated in figures 9 and 10.

The relatively low concentrations of dissolved solids near the ends of the aquifer result from infiltration of river water. The average dissolved-solids concentration in river water is much smaller than that in native ground water (table 3). Near the ends of the aquifer, differences in quality from one well to another reflect differences in the proportion of induced recharge to native ground water, as discussed in appendix F. Under maximum ground-water development, hardness of water from wells at the ends of the aquifer is unlikely to exceed 150 milligrams per liter and might approach the average hardness of water in the adjacent rivers, 75 to 95 milligrams per liter.

The observed increases in hardness and chloride may result from a variety of processes. Hardness in water is caused primarily by calcium and magnesium ions. Ground water in the Susquehanna River basin is usually hardest in areas of bright glacial deposits (Seaber and Hollyday, 1964) because such deposits contain abundant fragments of limestone and dolomite (carbonate rocks), which are especially rich sources of calcium and magnesium ions. Bright sand and gravel is widespread at the top of the Clinton Street-Ballpark aquifer (plate 6); however, drab sand and gravel predominates in the lower part (fig. 2). Chloride is rare in the natural earth materials that make up this aquifer, but is abundant at great depth in the bedrock and is readily leached from chemicals used for snow removal and from trash, sewage, and other products of human activities. Calcium chloride, sometimes used with common salt for snow removal, is a source of both hardness (due to the calcium) and chloride. Increased use of chemicals for snow removal during the 1950's and 1960's and increases in other activities accompanying greater urban development were probably a major factor responsible for increased hardness and chloride in ground water. Some wells were not pumped regularly until the 1960's, and shallow ground water that had elevated hardness because of contact with bright gravel and (or) with products of human activities may not have reached those wells until then. Also, slow seepage of relatively hard water from silt layers and salty water from deep in the bedrock into the aquifer may have accelerated somewhat during the 1960's, when water levels in the aquifer were lower than normal.

Leaky sewers are an important source of recharge and contamination of ground water in some urban areas (Kimmel, 1972) and might be the source of some of the nitrate and chloride in the Clinton Street-Ballpark aquifer, inasmuch as virtually all buildings above or near the aquifer are served by sanitary sewers. However, no significant increase in nitrate was recorded

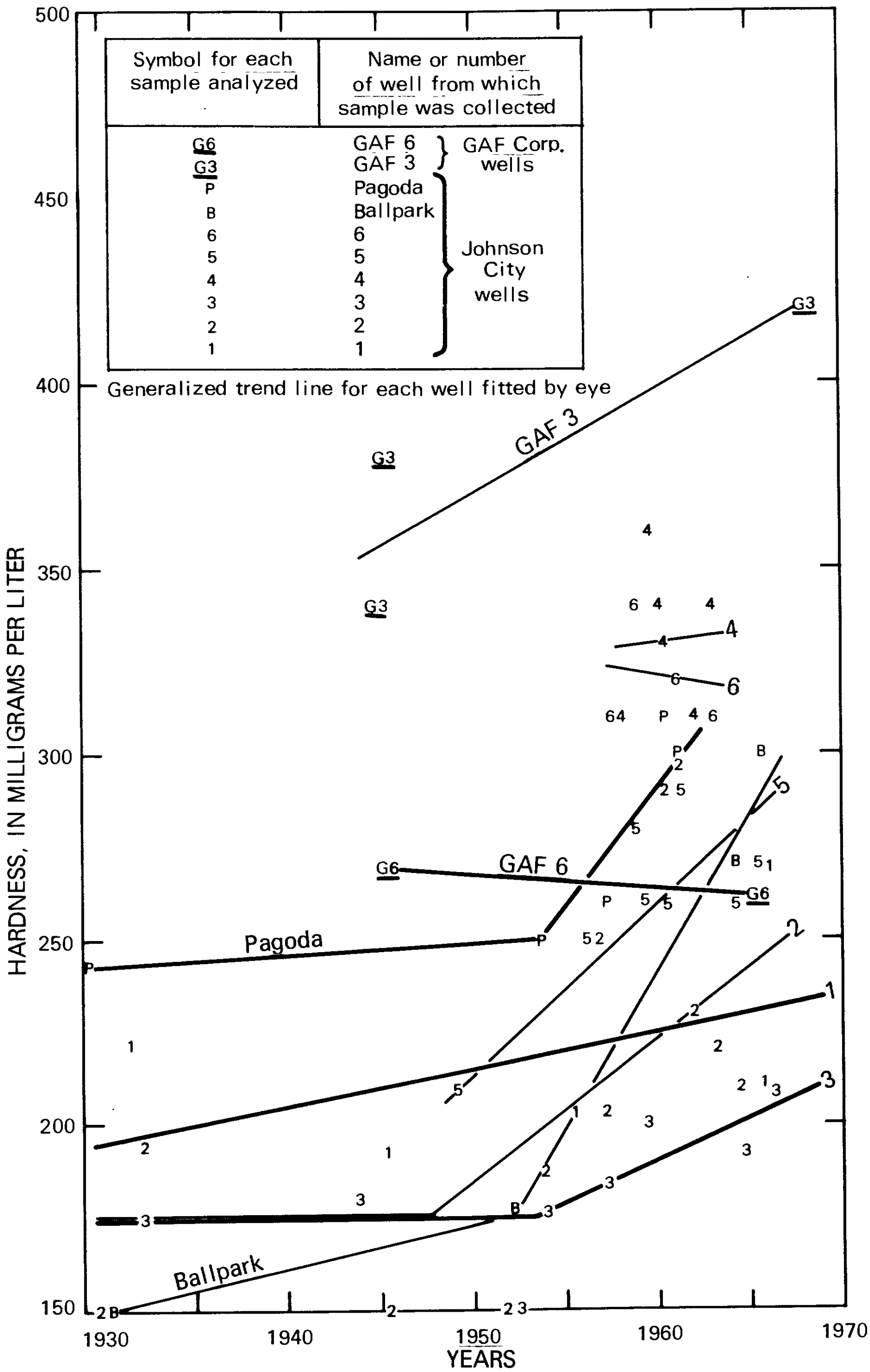


Figure 9.--Trends in hardness, 1929-69.

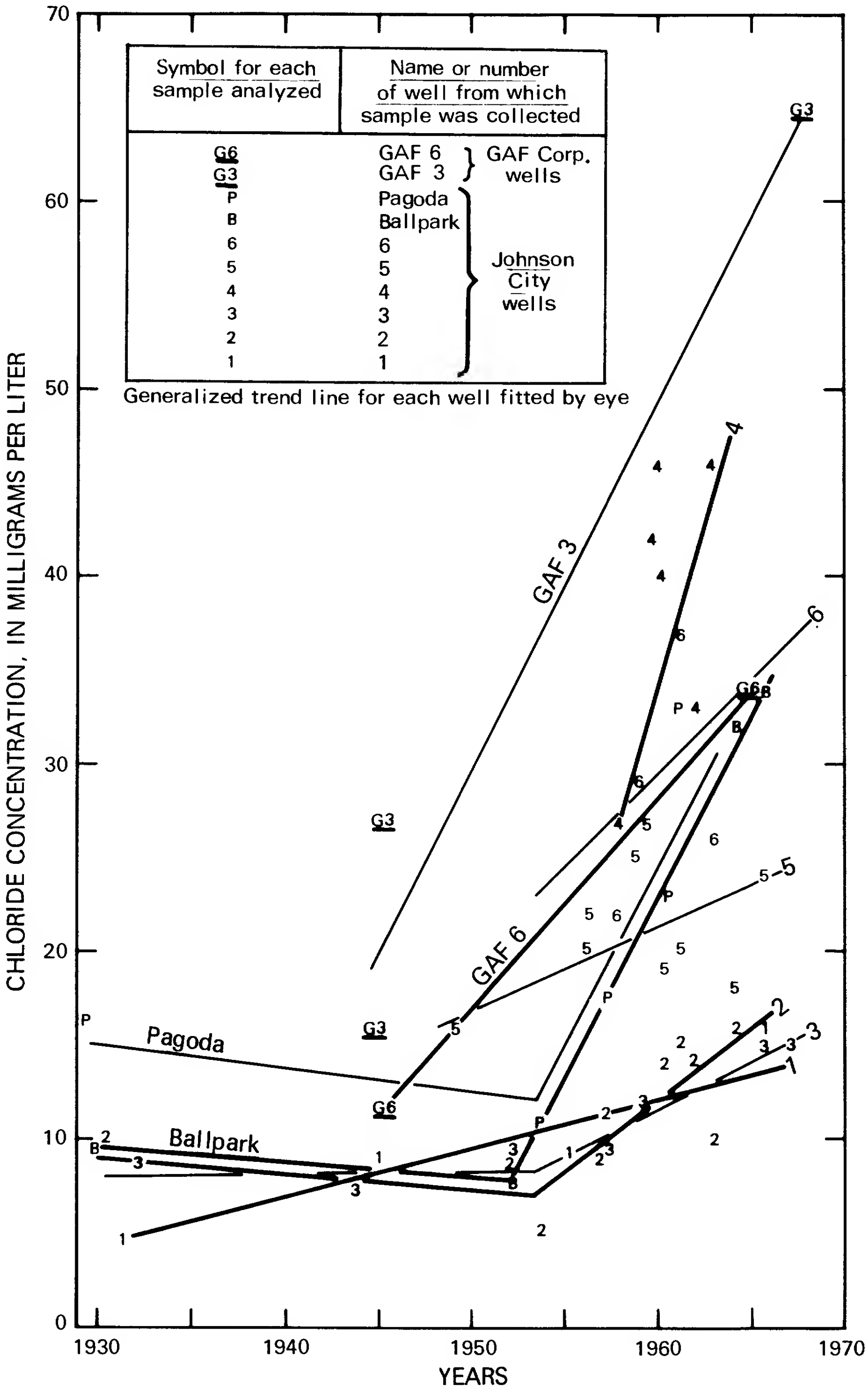


Figure 10.--Trends in chloride concentration, 1929-69.

Table 3.--Estimated average hardness, chloride concentration, and dissolved-solids concentration in water from several sources.

[All values in milligrams per liter.]

Constituent or property of water	Susquehanna River, approaching Binghamton 1/	Chenango River, approaching Binghamton 2/	Susquehanna River at Johnson City: North bank 3/	Susquehanna River at Johnson City: Entire river 4/	Ground water, Clinton Street-Ballpark aquifer 5/
Average hardness (as $CaCo_3$)	55	95	85	75	330
Average chloride concentration	4	4.5	5.5	6.0	35
Average dissolved-solids concentration (residue)	85	125	110	105	400-500

1/ Based on daily samples at Conklin, 1955 (Pauszek, 1959, p. 88), adjusted to represent long-term median flow (1931-60).

2/ Based on daily samples at Greene, 1957 (U.S. Geol. Survey, 1960) and miscellaneous samples elsewhere (Pauszek, 1959, p. 92); adjusted to represent long-term median flow (1931-60).

3/ Based on intermittent samples at Goudey Station in Johnson City, 1953-68 (unpub.), adjusted to represent long-term median flow (1931-60). Because of sewer outfalls upstream from Goudey Station and because the Chenango and Susquehanna Rivers do not mix thoroughly for several miles below their confluence (McDuffie, 1970), samples collected near the north bank at Goudey Station resemble Chenango River water more closely than Susquehanna River water.

4/ Estimated, assuming complete mixing of Chenango and Susquehanna Rivers and sewage.

5/ Based on latest samples analyzed through 1969 from wells not affected by induced recharge; dissolved solids estimated from measured hardness.

from 1950 to 1968 in any of the wells for which multiple analyses were available. Therefore, the observed increases in chloride and hardness during that period cannot be attributed to leaky sewers.

CONCLUSIONS RELATED TO AQUIFER MANAGEMENT

Responsibility for Management

The Clinton Street-Ballpark aquifer (fig. 1) is a geohydrologic unit, and maximum benefit can be obtained from it only if it is managed as a unit. Before 1969 there was little need for and no attempt made at joint planning by the various water users. If demands increase, however, withdrawals will be controlled increasingly by mutual interference if they are not controlled by mutual cooperation and foresight.

A water-supply survey for Broome County (Martin and Shumaker, 1968) called for establishing a water district to unite the several public water-supply systems in the Triple Cities area. If a regional water organization existed, management of aquifers that cross political boundaries could be simplified. However, because the interest of private industry in the Clinton Street-Ballpark aquifer seems to be as great as the public interest, judging from present well capacity and past pumpage, fair representation of both interests in management decisions will be necessary.

Whether management responsibility is vested in a formal organization or is exercised by informal cooperation, the problems and alternatives will be the same. The following sections were prepared as a guide to management.

Fundamental Considerations

The Clinton Street-Ballpark aquifer is already intensively developed, and in some years pumpage has exceeded recharge. Aquifer yield can be expanded beyond the 1968 level, but maximum yield consistent with optimum quality can be obtained only with careful management that takes into account two basic facts:

(1) Additional wells in the interior (central) part of the aquifer would not increase the total amount of water available; such wells could provide standby capacity during repair of other wells and could permit faster, more efficient withdrawal of water to meet peak demands. However, any water pumped from such wells would reduce the amount that could eventually be pumped from wells existing in 1968.

(2) Additional wells near the Chenango and Susquehanna Rivers at the ends of the aquifer would increase the total water supply by inducing river water to recharge the aquifer. At the east end, the rate at which most wells developed as of 1968 could induce recharge from the Chenango River was severely limited by distance and by layers of silt and clay within the aquifer. Temperature of the water from additional river-bank wells would fluctuate at least 6°C (10°F) and perhaps as much as 22°C (40°F) annually. The dissolved-solids content in water from such wells would be perhaps half that from wells in the interior of the aquifer, but bacteria from the river might reach some wells.

Alternative Management Options

The Clinton Street-Ballpark aquifer could be managed in different ways, depending on what objective is judged most important. Specifically, it could be managed to provide the largest possible yield of water for general municipal or industrial use--without concern for variations in temperature or chemical quality--or to provide low-temperature water suitable for air conditioning and year-round industrial cooling, in which case annual yield would not be as large. Either objective could be met in other ways; cooling could be effected by evaporation towers or refrigeration (at higher cost), and municipal demand could be supplied entirely from other aquifers or streams in the Triple Cities area. The Clinton Street-Ballpark aquifer is better suited than most to serve as a large reservoir of cool water, but it is also in the heart of the municipal service area. Either objective would call for construction of additional wells at the ends of the aquifer, but day-to-day management would differ, depending on which objective were chosen.

If maximum yield were the objective, wells near the ends of the aquifer could be pumped continuously to obtain as much induced recharge as possible. Wells in the interior of the aquifer could be pumped intermittently as needed, to meet maximum seasonal or daily demands (fig. 11). Peak demand normally occurs in midsummer. However, in late winter the rate of induced recharge could decline to 70 percent of the mean annual rate because of the relatively high viscosity of the cold river water; consequently, the need for supplementary pumpage from interior wells might be greatest in late winter. Aquifer yield in a dry year is presently estimated (table 2) to be at least 8,800 million gallons, or an average of 24 million gallons per day (91 million liters per day); information from future test wells and from measuring water levels prior to maximum development should permit more precise estimates of the maximum potential yield. Chemical quality and temperature of water in some parts of the distribution system might vary widely and often because these properties of induced recharge change seasonally and because withdrawal of native ground water from the interior of the aquifer, which differs in chemical character and often in temperature from induced recharge, would vary from day to day according to variation in demand.

A variation of the maximum yield option would pump wells in the interior of the aquifer to supply industrial uses that require water of constant low temperature or constant chemical quality and would pump end wells for other purposes (fig. 11). Some wells in each group could be reserved to meet peak demands. Recharge to the interior of the aquifer might be as little as 1,190 million gallons (4.5 billion liters) in a single dry year, or about 30 percent less than normal (table 2); during the prolonged drought of the 1960's, the cumulative deficiency might have been roughly 150 percent (fig. 5). However, approximately 1,320 million gallons (5 billion liters) of water stored underground could be withdrawn to balance a temporary deficiency in recharge by using only wells that existed in 1967, and a more uniform distribution of wells would make somewhat more stored water available (p. 14). Thus, if a prolonged drought were to recur, recharge plus storage in the interior of the aquifer could sustain an average withdrawal of at least 1,530 million gallons per year or 4.2 million gallons per day (16 million liters per day). (If the "high" estimate of recharge explained in appendix

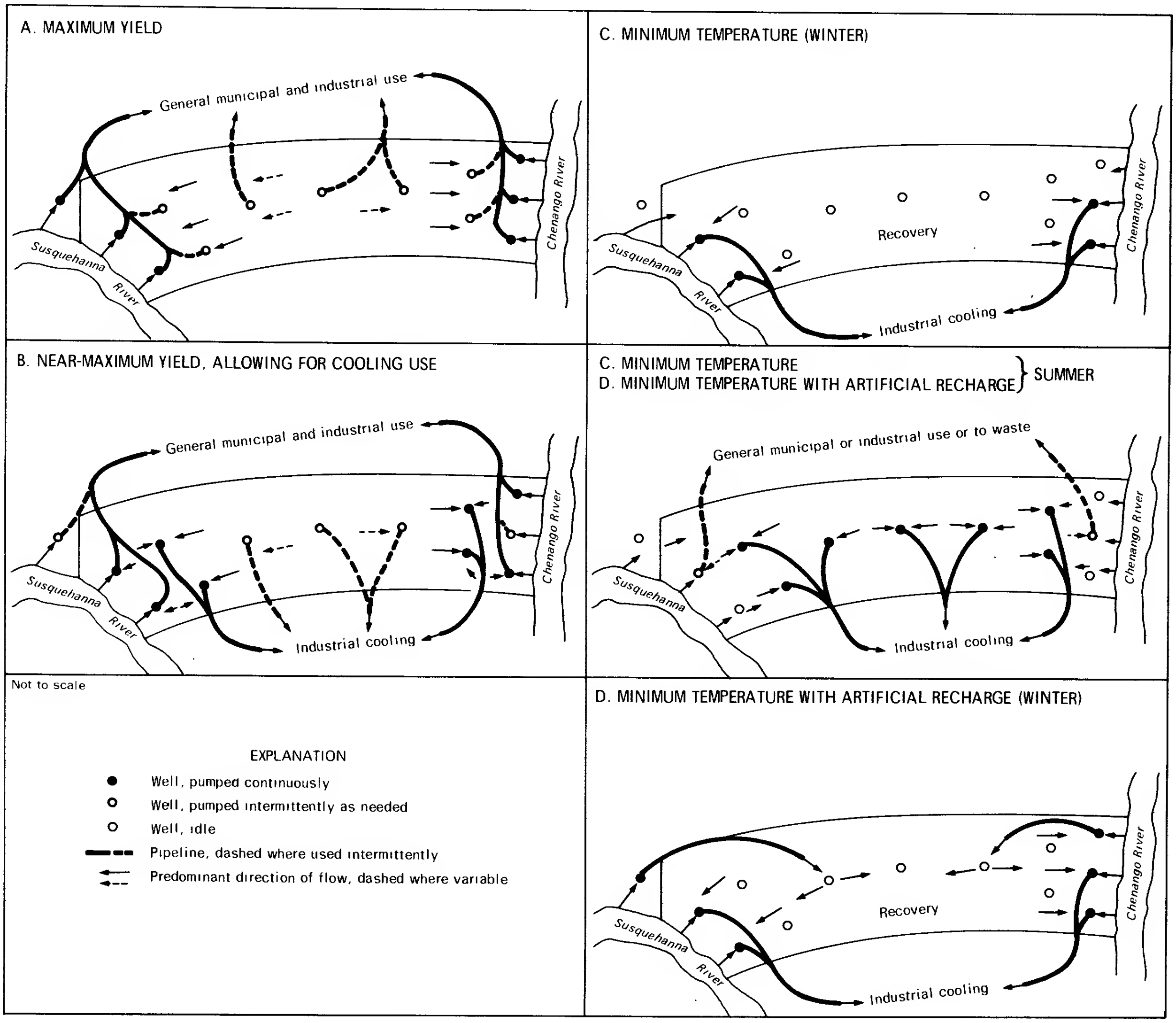


Figure 11.--Alternative options for aquifer management. Sketches illustrate only the general arrangement and purpose of ground-water withdrawals under each option, not the exact number of spacing of wells.

E is used, average withdrawal during droughts would become 4.9 million gallons per day, or 19 million liters per day). To insure that most of the cool native ground water of uniform quality remains available to interior wells rather than flowing to the wells near the ends of the aquifer that also induce recharge from the rivers, the outermost of the interior wells would have to be pumped regularly (fig. 11). Total aquifer yield under this option could be as great as under the previous option, 24 million gallons (91 million liters) per day, but only if variations in demand did not require some end wells to be shut down periodically, because this would cause induced infiltration to be less than the maximum potential.

If minimum temperature were the primary objective of aquifer management a more complex option would be called for. From mid-May to mid-October, river temperature is generally higher than ground-water temperature, which remains nearly constant at 12°C (53°F) in the interior of the aquifer. Consequently, to obtain the coolest water available, pumpage must be entirely from interior wells during the summer (fig. 11). Under drought conditions, recharge from precipitation from mid-May to mid-October could be virtually nil, in which case withdrawals should be limited to the amount of water in underground storage that would be replenished by recharge in subsequent winter seasons before another dry summer. In a drought year, recharge from precipitation should be at least 1,030 million gallons (3.9 billion liters) (table 2) and would occur chiefly from October to May; adding this to 7/12 of the annual recharge from Little Choconut Creek gives 1,120 million gallons (4.2 billion liters) as the net increase in storage over the winter. This amount of storage, plus the rest of the annual recharge from Little Choconut Creek, would permit a summer withdrawal of 7.9 million gallons per day (30 million liters per day). Clearly, summer withdrawal could exceed this rate, because 590 million gallons (2.2 billion liters) of the 1,700 million gallons (6.4 billion liters) estimated to be available in storage would not be touched. However, summer withdrawals could not be nearly as great as the estimated median annual recharge (table 2) of 1,510 plus 235 million gallons, or 11.5 million gallons per day (44 million liters per day), because this withdrawal rate would require all water available in storage to be used each summer, without leaving a reserve to sustain withdrawals during an extended drought. A summer withdrawal of about 1,300 million gallons, or 8.7 million gallons per day (33 million liters per day), is probably the maximum that could be sustained from the interior of the aquifer, if values of storage and recharge calculated in this report are approximately correct.

During the remaining 7 months, from mid-October to mid-May, cold water is available from wells at the ends of the aquifer (fig. 11). River temperature ranges from 0° to 12°C (32° to 53°F) during these months, and the temperature of water pumped could drop below 3°C (37°F) in late winter. The estimated induced infiltration potential (table 2) is ample to permit the proposed average summer withdrawal rate of 8.7 million gallons per day (33 million liters per day) to be continued throughout the winter.

If total pumping capacity of wells at the ends of the aquifer were to exceed the demand for cool water during the winter, the excess water could be conducted by pipeline to the interior of the aquifer and returned to the

ground without being used or heated. By this method of artificial recharge, aquifer storage could be replenished each winter, even in a dry year. Under such an option, the entire available aquifer storage (see section "Storage Available") plus recharge from Little Choconut Creek at the minimum rate (table 2) could be used each summer. This amounts to at least 1,770 million gallons, or an average of 11.8 million gallons per day (45 million liters per day) over the 150-day summer period. During the remaining 215 days of the year, induced river infiltration could sustain a demand of 11.8 million gallons per day (45 million liters per day) and also provide more than enough artificial recharge to replenish the deficit in storage in a dry year, as shown in the calculations below:

Induced recharge at ends of aquifer during winter period (mid-October to mid-May)		Recharge and storage in interior of aquifer	
16.5	million gallons per day potential for withdrawal from wells	1,700	million gallons storage available for use annually
-11.8	million gallons per day postulated demand	-1,120	million gallons natural recharge in winter of a dry year
4.7	million gallons per day surplus		
x 215	days		
Total 1,010	million gallons surplus in any year	Total 580	million gallons deficit in a dry year

Potential induced recharge at the ends of the aquifer during a typical winter is estimated to be 16.5 million gallons per day (62 million liters per day). This rate would be more than adequate, although it is about 20 percent less than the estimated annual rate in table 2 owing to the net effect of lower river temperature (which decreases infiltration because cold water has greater viscosity) and generally higher river stage (which partly compensates for the greater viscosity by increasing gradient toward the aquifer). Inasmuch as the surplus water piped to the interior of the aquifer in winter would be colder than 12°C (53°F), withdrawals during the summer would be slightly colder as well as larger than would have been possible without this artificial recharge.

Two aspects of the minimum-temperature option would require special attention by water managers. First, some undesirably warm water would be drawn into the aquifer by induced infiltration during the summer, although at a reduced rate because the wells nearest the ends of the aquifer would be shut down. This effect could be minimized by withdrawing water chiefly

from the most interior part of the aquifer in early summer and by pumping the shallowest wells at the ends of the aquifer to waste, or for some use not sensitive to temperature, during the latter part of the summer; pumping the shallowest wells would intercept and remove some of the warmest water. However, under maximum development it would be impossible to avoid having water considerably warmer than 12°C (53°F) near most end wells by the end of the summer; hence, these wells would have to be pumped to waste for a few days or weeks until cooler induced recharge reached them and lowered the temperature of pumped water to acceptable levels. The second point to remember is that each spring, at the time of changeover from pumping end wells to interior wells, the dissolved-solids content of water in the distribution system would increase markedly; it would decrease correspondingly each fall when the end wells were again pumped.

The alternative management options described in the preceding paragraphs illustrate the aquifer's potential for meeting the type of demands imposed on it during the 1960's. Compromises or other alternatives would also be possible. Presentation of a detailed plan for use of the aquifer, which would involve considering future demands and other sources of water available to the Triple Cities, is beyond the scope of this report.

Artificial Recharge

The concept of supplementing natural aquifer recharge by artificial means has become widely accepted, especially in arid regions. In general, the idea is to store temporary streamflow surpluses underground, thereby increasing aquifer yield. Recirculation of "used" ground water is practiced in some places on Long Island; for example, ground water pumped for cooling is returned to the ground (Johnson, 1955), many homes dispose of wastewater through leach fields, and consideration is being given to injecting treated municipal sewage into aquifers (Cohen and others, 1968, p. 100). Many techniques of artificial recharge have been tested and are described in the extensive literature on the subject (Todd, 1959, 1963; Bauman, 1965; Signor and others, 1970). At least three techniques may be applicable to the Clinton Street-Ballpark aquifer:

(1) Recharge the interior of the aquifer with surplus induced infiltration pumped from the ends of the aquifer. Potential aquifer yield could be increased by 3 million gallons per day (11 million liters per day) by this technique under certain conditions, as described in the preceding section. A similar technique is already being practiced on a small scale in the area; several GAF production wells are pumped continuously, and, whenever a temporary decline in demand causes the company's storage tank to overflow, the overflow is returned to the ground through a group of abandoned wells at the main GAF plant on Charles Street. An average of 120 million gallons per year (450 million liters per year) has been returned since 1958 (appendix B). More wells could be devoted to artificial recharge, or production wells and plumbing systems could be designed to alternately discharge and receive water (McMillian and Hauser, 1969; Bauman, 1965, p. 276), or water could be returned economically through shallow dug wells or pits in permeable gravel above the water table. Inasmuch as the well water to be used for recharge would already have been filtered through the aquifer, clogging

of recharge wells would probably not be a major problem. Although double pumping would raise the unit cost of water, electrical demand charges for pump startups could be minimized by operating wells at the ends of the aquifer continuously and recharging during off-peak hours. With prudent aquifer management, little or none of the artificial recharge water would be lost. In unusually wet years, there would probably be little need for artificial recharge.

(2) Increase recharge from Little Choconut Creek. Little Choconut Creek has been a significant source of recharge to the Clinton Street-Ballpark aquifer, but, over the years, progressively larger reaches of this and other local creeks have been encased in concrete, consequently decreasing recharge from the creek. Continuation of this process would eventually eliminate the creek as a source of recharge. However, recharge could be increased in at least two ways:

(a) Deepen present channels within the area of surficial silt (plate 6), which generally extends only a few feet below the streambed, and remove oily streambed mud and trash along with the silt. To minimize deposition of new mud, the deep pools thus created could be backfilled with clean gravel. The resulting increase in infiltration rate might be on the order of 0.6 cubic feet per second per 1,000 feet of treated channel (17 liters per second per 300 meters).

(b) Excavate multiple channels or small basins through the surficial silt into the underlying gravel. Minor regulation of the flood-control reservoirs upstream could help keep the basins or channels full for substantial periods of the year.

Any increase in streambed permeability, whether in new or old channels, would probably increase the distance to which bacteria infiltrate into the aquifer. Sanitary quality of water from Little Choconut Creek has not been studied, but several domestic sewage outfalls were emptying into the creek upstream from the aquifer as of 1968.

(3) Increase the infiltration of local storm runoff. As urban business districts expand, ever-larger areas are made impervious by buildings and pavement. So far, recharge to the Clinton Street-Ballpark aquifer does not seem to have been drastically reduced by this process; nevertheless, nearly all water diverted to the rivers by storm sewers in areas of stratified drift represents recharge lost to the aquifer. This loss could be minimized if runoff from large buildings, lots, and some streets were conducted to perforated storm sewers or recharge basins, as is done in Long Island to augment ground-water supplies (Welsch, 1949, 1956; Seaburn, 1970; Seaburn and Aronson, 1974). However, unless runoff from major streets and highways is diverted to the rivers, the large amounts of salt dissolved during the winter would be added to the ground water.

Observation Wells

Changes in the amount of water stored in the Clinton Street-Ballpark aquifer may be monitored by periodically measuring the water level in many idle wells. Annual measurements around October 1 (when water levels are lowest) should be adequate to define long-term trends. With maximum aquifer development, more frequent measurement would be needed. Water levels in

many operating wells are measured monthly or weekly to determine well efficiency and to avoid excessive drawdown, but measurements in such wells do not provide a reliable index of changes in storage. Likewise, frequent observations in one or two key wells are useful but may reflect local conditions not typical of the entire aquifer.

The locations of all wells known to the Geological Survey in which water levels could be measured as of 1971 are shown in plate 1. The altitude and a description of the measuring point of each well are given in appendix G.

Although the Clinton Street-Ballpark aquifer is penetrated by more observation wells than any other aquifer of comparable size in the Susquehanna River basin, a few parts of the aquifer are rather sparsely represented. Additional observation wells in two localities would be especially useful:
(1) on islands in the Susquehanna River and along the riverbank at the west end of the aquifer, screened at different depths, to define the cone of depression under the river and the potential for induced recharge; and
(2) at the city line, to better define flow directions near the groundwater divide. The Pagoda well, an abandoned production well listed in appendix G, would be ideal for the latter purpose but was not accessible as of 1968.

Aquifer Modeling

Techniques are now available for constructing models of aquifers, both electric-analog models consisting of small electrical components wired on a board, and digital models consisting of an electronic-computer program. A model is usually tested by comparison with a known history of pumpage and water-level changes and is adjusted until simulation of the historical pumpage results in simulated water-level changes corresponding in magnitude and extent to those actually observed. The model can then be used to predict the response of the aquifer to any postulated management regime. The techniques and results of constructing both types of model for a small sand-and-gravel aquifer are described in Pinder and Bredehoeft (1968).

Water managers often find an aquifer model useful in selecting the number, spacing, design specifications, and operating schedules of new wells when further development of an aquifer is under consideration. This report provides much of the information that would be needed for initial construction and testing of a model of the Clinton Street-Ballpark aquifer.

APPLICATION OF THIS REPORT TO OTHER AREAS

As pointed out earlier, the most productive aquifers in the New York part of the Susquehanna River basin are bodies of sand and gravel that are interbedded with silt and clay in varying proportions and are confined to major valleys. Large streams flow through nearly all these valleys, and water can move fairly easily between stream and aquifer except where silt

or clay layers intervene. In such valleys, some of the constraints on aquifer yield and management would differ substantially from those described in this report.

A few other valleys in the Susquehanna River basin contain sand-and-gravel aquifers that are separated from the major stream, like the Clinton Street-Ballpark aquifer. The most similar example lies a few miles north of Binghamton, where a broad valley runs west from the Chenango River through Kattellville, then south to Chenango Bridge (fig. 1). Other examples are at Big Flats in Chemung County, at Cortland in Cortland County, and at Sherburne in Chenango County. The various plans discussed in the section "Alternative Management Options" could also be applied to aquifers in these localities, with the objective of obtaining either maximum yield or minimum temperature. If the objective were maximum yield with minimum depletion of river flow during critical low-flow periods--an especially attractive goal for some of these localities, where river flow is less than it is near Binghamton--the likely approach to development would be nearly identical to the minimum-temperature option, except that the decision to shift from use of riverbank wells to use of wells in the interior of the aquifer would be based on amount of river flow rather than river temperature.

The recharge rates from precipitation presented in this report apply to surficial sand-and-gravel aquifers regardless of whether major streams are present. Approximate adjustments for differences in annual rainfall may be made using figure E2. Rates of recharge from Little Choconut Creek are probably not applicable elsewhere because preliminary analysis of measurements from other tributary streams suggests that most of them have higher rates of infiltration per unit area or unit length of stream channel than Little Choconut Creek.

REFERENCES CITED

Barksdale, H. C., O'Bryan, Deric, and Schneider, W. J., 1966, Effect of drought on water resources in the northeast: U.S. Geol. Survey Hydrol. Inv. Atlas HA-243.

Bauman, Paul, 1965, Technical development in ground-water recharge, in Chow, V. T., ed., Advances in hydroscience, v. 2: New York, Academic Press, p. 209.

Bedinger, M. S., 1961, Relation between median grain size and permeability in the Arkansas River valley, Arkansas: U.S. Geol. Survey Prof. Paper 424-C, p. C31-C32.

Bentall, Ray, ed., 1963, Shortcuts and special problems in aquifer tests: U.S. Geol. Survey Water-Supply Paper 1545-C., 117 p.

Bouwer, Herman, 1964, Unsaturated flow in ground-water hydraulics: Am. Soc. Civil Engineers Proc., Jour. Hydraulics Div., v. 90, no. HY 5, p. 121-139.

__________, 1965, Theoretical aspects of seepage from open channels: Am. Soc. Civil Engineers Proc., Jour. Hydraulics Div., v. 91, no. HY 3, p. 37-59.

Brown, R. H., and Ferris, J. G., 1946, Progress report on ground-water resources of the southwestern part of Broome County, New York: New York Water Power and Control Comm. Bull. GW-15.

Cadwell, D. H., 1973, Glacial geology of the Chenango River valley near Binghamton, N.Y., in Coates, D. R., ed, Glacial geology of the Binghamton-western Catskill region: State Univ. New York at Binghamton, Pub. in Geomorphology, contr. 3, p. 31-39.

Coates, D. R., 1966, Discussion of K. M. Clayton "Glacial erosion in the Finger Lakes region (New York, U.S.A.)": Berlin, Zeitschr. Geomorphologie, neue Folge, Bd. 10, Heft 4, S. 469, [new ser. v. 10, no. 4, p. 469].

Cohen, Philip, Franke, O. L., and Foxworthy, B. L., 1968, An atlas of Long Island's water resources: New York Water Resources Comm. Bull. 62, 117 p.

Denny, C. S., and Lyford, W. H., 1963, Surficial geology and soils of the Elmira-Williamsport region, New York-Pennsylvania: U.S. Geol. Survey Prof. Paper 379, 60 p.

Ferris, J. G., Knowles, D. B., Brown, R. H., and Stallman, R. W., 1962, Theory of aquifer tests: U.S. Geol. Survey Water-Supply Paper 1536-E, 174 p.

Heath, R. C., 1964, Ground water in New York: New York State Dept. Conservation Bull. GW-51.

Hollyday, E. F., 1969, An appraisal of the ground-water resources of the Susquehanna River basin in New York State: U.S. Geol. Survey open-file rept., 52 p.

Hunt, O. P., 1967, Duration curves and low-flow frequency curves of streamflow in the Susquehanna River basin: New York Water Resources Comm. Bull. 60, 52 p.

Hurr, R. T., 1966, A new approach for estimating transmissibility from specific capacity: Water Resources Research, v. 2, no. 4, p. 657-664.

Johnson, A. H., 1955, Conservation of ground water on Long Island: Am. Water Works Assoc. Jour., v. 47, no. 4, p. 348-354.

Kimmel, G. E., 1972, Nitrogen content of ground water in Kings County, Long Island, N.Y.: U.S. Geol. Survey Prof. Paper 800-D, p. D199-D203.

Ku, H. F. H., Randall, A. D., and MacNish, R. D., 1975, Streamflow in the New York part of the Susquehanna River basin: New York State Dept. Environmental Conserv. Bull. 71, 130 p.

Leopold, L. B. and Maddock, Jr., Thomas, 1953, The hydraulic geometry of stream channels and some physiographic implications: U.S. Geol. Survey Prof. Paper 252, 57 p.

Lohman, S. W., 1972, Ground-water hydraulics: U.S. Geol. Survey Prof. Paper 708, 70 p.

Martin, R. J. and Shumaker, V. O., 1968, Development of water facilities in Broome County, 1967-2017: CPWS-8, prepared for Broome County Board of Supervisors.

McDuffie, Bruce, 1970, A brief chemical study of the Chenango and Susquehanna Rivers, with some data on lake samples: mimeo. rept. under contract between Broome County and Research Foundation of the State University of New York.

McMillian, L. G. and Hauser, V. L., 1969, Field evaluation of potential pollution from ground-water recharge: Water Well Jour., v. 23, no. 8, p. 22.

Meinzer, O. E., and Stearns, N. D., 1929, A study of ground water in the Pomperaug basin, Connecticut: U.S. Geol. Survey Water-Supply Paper 597-B.

Moore, J. E. and Jenkins, C. T., 1966, An evaluation of the effect of ground-water pumpage on the infiltration rate of a semipervious streambed: Water Resources Research, v. 2, no. 4, p. 691-696.

Moss, J. H., and Ritter, D. F., 1962, New evidence regarding the Binghamton substage in the region between the Finger Lakes and the Catskills: Am. Jour. Sci., v. 260, p. 81.

Olmsted, F. H., and Hely, A. G., 1962, Relation between ground water and surface water in Brandywine Creek basin, Pennsylvania: U.S. Geol. Survey Prof. Paper 417-A., 21 p.

Parker, G. G., Hely, A. G., Keighton, W. B., Olmsted, F. H., and others, 1964, Water resources of the Delaware River basin: U.S. Geol. Survey Prof. Paper 381, 200 p.

Pauszek, F. H., 1959, Chemical quality of surface waters in the Allegheny, Genesee, and Susquehanna river basins, New York, 1953-56: New York State Dept. Commerce Bull. 3.

Pinder, G. F. and Bredehoeft, J. D., 1968, Application of the digital computer for aquifer evaluation: Water Resources Research, v. 4, no. 5, p. 1069-1093.

Pluhowski, E. J., and Kantrowitz, I. H., 1964, Hydrology of the Babylon-Islip area, Suffolk County, Long Island, New York: U.S. Geol. Survey Water-Supply Paper 1768, 119 p.

Randall, A. D., 1964, Geology and ground water in the Farmington-Granby area, Connecticut: U.S. Geol. Survey Water-Supply Paper 1661, 129 p.

__________, A. D., 1970, Movement of bacteria from river to municipal well: a case history: Jour. Am. Water Works Assoc., v. 62, no. 11, p. 716-720.

__________, 1972, Records of wells and test borings in the Susquehanna River basin, New York: N.Y. State Dept. Environmental Conservation Bull. 69, 92 p.

__________, in press, A contribution to the Pleistocene stratigraphy of the Susquehanna River basin: N.Y. State Education Dept., Empire State Geogram.

Randall, A. D., Thomas, M. P., Thomas, C. E., and Baker, John, 1966, Water resources inventory of Connecticut, part 1, Quinebaug River basin: Conn. Water Resources Bull. 8, 102 p.

Rorabaugh, M. I., 1956, Ground-water resources of the northeastern part of the Louisville area, Kentucky: U.S. Geol. Survey Water-Supply Paper 1360-B., 168 p.

Seaber, Paul and Hollyday, E. F., 1964, Hardness of ground water in the glacial deposits of the Susquehanna River basin, New York and Pennsylvania: Geol. Soc. America, Program 1964 Ann. Mtg. [abs. only] and U.S. Geol. Survey open-file rept.

Seaburn, G. E., 1970, Preliminary results of hydrologic studies at two recharge basins on Long Island, New York: U.S. Geol. Survey Prof. Paper 627-C, p. C1-C17.

Seaburn, G. E. and Aronson, D. A., 1974, Influence of recharge basins on the hydrology of Nassau and Suffolk Counties, Long Island, New York: U.S. Geol. Survey Water-Supply Paper 2031, 66 p.

Signor, D. C., Growitz, D. J., and Kam, William, 1970, Annotated bibliography on artificial recharge of ground water 1955-67: U.S. Geol. Survey Water-Supply Paper 1990, 141 p.

Todd, D. K., 1959, Annotated bibliography on artificial recharge of ground water through 1954: U.S. Geol. Survey Water-Supply Paper 1477, 115 p.

________, 1963, Ground-water hydrology: New York, John Wiley & Sons, Inc., 336 p.

Turcan, A. N., Jr., 1963, Estimating the specific capacity of a well: U.S. Geol. Survey Prof. Paper 450-E, p. E145-E149.

Turner, J. F., Jr., 1966, Evaporation study in a humid region, Lake Michie, North Carolina: U.S. Geol. Survey Prof. Paper 272-G, p. G137-G150.

U.S. Geological Survey, 1951-57, Water levels and artesian pressures in wells in northeastern states: U.S. Geol. Survey Water-Supply Papers 1096, 1126, 1156, 1165, 1191, 1221, 1265, 1321, 1404.

________, 1960-70, Ground-water levels in the United States, northeastern states: Water-Supply Papers 1537, 1782, 1977.

________, 1960, Quality of surface waters of the United States, 1957, parts 1-4: U.S. Geol. Survey Water-Supply Paper 1520, 641 p.

________, 1970, Water resources data for New York, 1969, part 1, Surface water records: U.S. Geol. Survey open file rept. (issued annually). [Available from U.S. Geological Survey, Albany, N.Y., 12201.]

U.S. Public Health Service, 1962, Drinking water standards, 1962: U.S. Public Health Service Pub. 956, 61 p.

U.S. Weather Bureau, 1965, Climatological data New York: U.S. Weather Bur., Climatological data, v. 76, no. 13, p. 230-241.

Walton, W. C., 1962, Selected analytical methods for well and aquifer evaluation: Illinois State Water Survey Bull. 49.

____________, 1963, Estimating the infiltration rate of a streambed by aquifer-test analysis: Berkeley, Calif., Internat. Assoc. Scientific Hydrology, Pub. 63, p. 409-420.

Walton, W. C., Hills, D. L., and Grundeen, G. M., 1967, Recharge from induced streambed infiltration under varying ground-water-level and stream-stage conditions: Univ. Minnesota Graduate School Water Resources Center, Bull. 6.

Welsch, W. F., 1949, Conservation of ground-water resources, Nassau County, New York: Water Works Eng., v. 102, no. 8, p. 708.

____________, 1956, Conservation of ground water; practices used on Long Island: Water and Sewage Works, v. 103, no. 10, p. 469.

Wolman, M. G., 1955, The natural channel of Brandywine Creek, Pennsylvania: U.S. Geol. Survey Prof. Paper 271, 56 p.

APPENDICES A - G

APPENDIX A: COMPUTATION OF TRANSMISSIVITY

Three methods were used to obtain the transmissivity values from which plate 3 was drawn.

(1) Pumping tests. Data from six pumping tests, chiefly driller's tests, were analyzed by time- and distance-drawdown methods (Walton,1962). In each test, the transmissivity values that could be calculated from different observation wells differed widely owing to local inhomogeneities or boundaries within the aquifer.

(2) Specific capacity. Transmissivity was calculated from reported specific capacity for 31 wells according to techniques described by Hurr (1966), Walton (1962), and Theis and others (in Bentall, 1963). These techniques consider well radius and pumping time but require an estimate of the storage coefficient. A correction for the geometric aspects of partial penetration (in an isotropic aquifer) was made by reversing a procedure described by Turcan (1963). This correction more than doubled many transmissivities. Dewatering was corrected for where significant (Walton, 1962, p. 7). Well loss was neglected for most wells because tests were made shortly after development or redevelopment and because drawdowns were only 1 to 20 feet (0.3 to 6 meters); aquifer anisotropy, boundaries, and well development also were neglected. All but the last of these factors would generally have raised estimates of transmissivity had it been possible to allow for them.

(3) Well logs. Hydraulic conductivity of unconsolidated sediment generally increases with grain size and with improved sorting. If the hydraulic conductivity of each saturated lithologic unit penetrated by a well can be determined, transmissivity may be calculated by multiplying the thickness of each unit by its hydraulic conductivity and summing the products. Bedinger (1961) worked out an average relationship between median grain size and hydraulic conductivity for fluvial sediments of the Arkansas River valley. From his results and data from Connecticut (Randall and others, 1966, p. 51), hydraulic conductivity was estimated for each lithologic unit penetrated by each well whose specific capacity was known. Transmissivities thus computed from well logs were compared with transmissivities computed from specific capacity, and estimates of hydraulic conductivity for some materials were revised. Final estimates are given in table A1. Although for most wells, results of the two methods differ by a factor less than 3, correlation is weak, probably owing to difficulty in estimating the hydraulic conductivity of gravel. Gravel layers observed in excavations range from openwork pebble gravel whose hydraulic conductivity must be greater than 1,300 feet per day (400 meters per day) to silt-bound sandy gravel whose hydraulic conductivity is nearly nil. A good log would indicate a notably silty gravel, but thin openwork layers might be overlooked or imprecisely described.

The hydraulic conductivity values in table A1 were applied to the logs of 96 wells and test borings in or near the Clinton Street-Ballpark aquifer to calculate transmissivity. Several logs that did not reach bedrock were extended 10 to 60 feet (3 to 18 meters) on the basis of bedrock contours (plate 2), cross sections (fig. 3), and nearby logs.

Appendix A

Table A1.--Estimated hydraulic conductivities of aquifer materials

Material	Hydraulic conductivity (feet per day)
Sample-study logs by geologist or soils technician	
Gravel, fine to coarse, little or no sand or silt	2,700
Gravel, some sand, trace silt	2,000
Gravel and sand, trace silt, loose	1,300
Sand, coarse to very coarse, pebbly, clean, water-yielding	1,300
Sand, medium to very coarse, and gravel, clean, water-yielding	1,300
Sand, medium to very coarse, and gravel, slightly silty, water-yielding	1,000
Sand, medium to very coarse, clean, water-yielding (no gravel)	1,000
Sand, medium to very coarse, slightly silty	700
Sand, fine to coarse, pebbly, clean	700
Sand, some gravel	700
Sand, fine to coarse, pebbly, slightly silty	500
Sand, fine, some gravel	270
Sand and gravel, moderately silty	130
Silty sand and gravel	50
Driller's log or equivalent	
Gravel, coarse, water-yielding or screened	1,300
Gravel, coarse, not screened or described as water-yielding	1,000
Coarse sand and gravel, water-yielding or screened	1,000
Gravel, medium, not screened or described as water-yielding	800
Sand and gravel, or gravel, water-yielding or screened (no size modifiers, presumed to be fine-to-coarse sand)	700
Sand and gravel, or gravel, not screened or described as water-yielding	500
Fine sand and gravel, sand and gravel with some fine sand; not screened	270
Sand and gravel, some silt or clay	130
Gravel and silt	25
Gravel and clay	15
All logs	
Medium sand, fine sand, very fine sand, sand (presumed medium); clean to silty: transmissivity selected from graph (Randall and others, 1966, p. 51)	1 to 270
Silt, clay, hardpan	.1

All transmissivity calculations were adjusted to water levels thought to be the maximum obtainable under developed conditions: 825-830 feet (252 meters) in altitude near the city line and the Chenango River, 820 feet (250 meters) over much of the aquifer, and 810-815 feet (248 meters) near the Susquehanna River. Final values are shown in plate 3; transmissivity contours are based on these point values and also on interpretation of geologic sections (fig. 3), contours on the base of the aquifer (plate 2), and water-level gradients (plate 8). A predominance of east-west belts of similar transmissivity was noted in the eastern part of the aquifer; consequently a similar pattern was drawn in the western part, although the data were less consistent.

Transmissivity was also calculated from parts of the flow net for October 1967 (plate 8) that closely encircle major pumping centers (Lohman, 1972, p. 46). Values thus computed represent steady-state flow across a perimeter a few thousand feet long, rather than point data such as used to prepare plate 3. Allowing for the lateral changes in transmissivity suggested by plate 3, values computed from the flow net are one-half to possibly one-fifth as large as those computed from the point data. Water-level gradients in plate 8 are imprecise but could not reasonably be reduced by factors of 2 to 5. Perhaps fine-grained layers are distributed within the aquifer in such a way as to commonly form partial barriers between permeable gravel units (fig. 3), thereby limiting overall aquifer transmissivity while permitting high values at most individual points. Accordingly, transmissivity values from plate 3 probably should be reduced by about one-half before use in preliminary aquifer modeling.

* * *

Table B1.--Pumpage and artificial recharge, Clinton Street-Ballpark aquifer, September 1958 to October 1968

[In millions of gallons. Horizontal lines in first column indicate groups of wells whose pumpage was combined when flow lines were drawn (pl. 7, 8) and recharge was computed (table E1).]

Point of withdrawal or recharge 1/	Source of pumpage data 2/	Pumpage or (if preceded by negative sign) recharge									
		1958 3/	1959	1960	1961	1962	1963	1964	1965	1966	1967 4/
Int. Business Machines Country Club wells	E2	1	11	11	29	29	29	29	29	29	27
Johnson City wells 1-3											
Total pumpage	M	519	1784	1343	1344	1348	1318	1295	1325	1377	1068
Pumpage derived from aquifer	E3	270	933	501	508	510	498	490	501	520	405
Johnson City well 5	M	--	40	75	136	39	46	153	109	83	113
Johnson City wells											
Well 4	M	--	--	130	140	76	52	--	--	--	--
Well 6	M	--	--	255	446	599	486	55	77	23	43
Ballpark well	H	7	600	334	34	80	151	750	765	789	537
Pagoda well	H	64	--	164	113	92	66	--	--	--	--
GAF Camera Plant well	E	1	3	3	3	3	3	3	3	3	2
Fairbanks Co. well	E1	--	4	4	4	4	4	4	4	4	4
GAF Charles Street wells											
Pumpage, wells 3, 5	O	151	470	480	625	479	435	435	469	418	327
Artificial recharge	O	5/ -72	-250	-240	-193	-68	-110	-55	-28	-60	-54
Titchener Co. well 6/	E1	8	36	36	36	36	38	38	36	36	28
GAF well 10	O	--	183	224	92	197	216	198	36	92	2
GAF well 6	O	5/ 20	90	43	62	120	115	105	85	87	124
GAF well 9	O	5/ 165	643	499	401	383	496	377	376	397	345
GAF well 8	O	5/ 136	443	417	570	390	320	419	248	220	250
Cutler Ice. Co. well	E	6	26	26	26	26	26	26	26	26	20
GAF well 7	O	5/ 125	454	490	415	488	529	450	506	479	262

1/ Wells arranged from west to east
2/ Symbols defined as follows:
- E estimated
- E1 estimated from pump rated capacity and owner's recollection of hours normally operated
- E2 estimated from pump rated capacity and owner's recollection of hours normally operated, multiplied by 0.8 to allow for recharge from irrigation return water and for days not used because of rain
- E3 estimated as 52 percent of total pumpage for 1958-59; 37 percent thereafter; based on flow lines in plates 7 and 8. Remainder of pumpage originates in or beyond Susquehanna River

2/ (continued)
- H estimated from test of pump capacity and from owner's record of hours operated
- M measured by propeller meter
- O measured by pressure drop across constriction in pipeline

3/ September 25 to December 31
4/ January 1 to October 6
5/ Estimated as 25 percent of measured annual total
6/ Half of pumpage combined with GAF wells 3 and 5, half with GAF wells 6 and 10 for computation of recharge

APPENDIX C: COMPUTATION OF AVAILABLE STORAGE

One approach to estimating the storage capacity of the Clinton Street-Ballpark aquifer is to look at past performance. The long-term net decline in water level from September 25, 1958 to October 6, 1967 (plates 7, 8) dewatered about 720 million cubic feet (20 million cubic meters) of the aquifer. Each spring, recharge exceeds pumpage, and therefore storage is partly replenished, but during the summer, storage is depleted again (fig. 6). Comparison of water-level contours for April and October 1967 showed that about 385 million cubic feet (11 million cubic meters) of the aquifer were dewatered that summer. A comparable seasonal decline in water level must have taken place in 1958 prior to the measurements on September 25. Therefore, the total of observed seasonal plus long-term reduction in saturated volume of aquifer materials was about 1,105 million cubic feet (31 million cubic meters). If the aquifer materials contain 16 percent pore space from which water will drain by gravity--a reasonable estimate if dewatering were to take place over a single summer season--then usable storage is about 177 million cubic feet or 1,320 million gallons (5 billion liters) of water. This figure seems realistic for the 1967 well spacing and for the year-round pumpage pattern that has prevailed for many years. However, it is conservative as an index of maximum available storage in that:

(a) the large cones of depression that existed in April 1967 and September 1958 (starting points for the two computations) would be available to store additional water if the wells were not pumped year-round;

(b) a few new wells at strategic locations could substantially increase the amount of water that could be withdrawn in a single season.

Another approach is to consider gross aquifer volume. By subtracting the altitude of the base of the aquifer (plate 2) from the maximum water levels used in estimating transmissivity (appendix A), we obtain a gross aquifer volume of about 5,700 million cubic feet (160 million cubic meters). Applying a gravity yield of 16 percent, we obtain 912 million cubic feet or 6,800 million gallons (26 billion liters) of water. Because of the geometry of the cone of depression around a production well, it is usually impossible to withdraw more than about one-third of this stored water in one season without having an expensively large number of closely spaced wells. Therefore, we may estimate usable storage as nearly 2,300 million gallons (8.7 billion liters) of water.

From the results of these two approaches, the maximum available storage was assumed to be 1,700 million gallons (6.4 billion liters) for purposes of calculations in this report.

* * *

APPENDIX D: RECHARGE FROM LITTLE CHOCONUT CREEK

The rate at which water infiltrates from Little Choconut Creek and its tributaries to the Clinton Street-Ballpark aquifer may be determined by measuring flow in each stream where it begins to cross the aquifer (that is, where it leaves till and bedrock or issues from a culvert) and comparing the total thus measured with the flow remaining where Little Choconut Creek enters the Susquehanna River. Flow was measured at numerous sites (plate 5) on 16 dates from June 1966 through November 1968, and the raw measurements have been published (U.S. Geological Survey, 1970). All 16 sets of measurements showed a net loss from the stream system, but the precision with which loss was determined is rather low. On the following pages, sources of error and adjustments made to the raw data are described, adjusted loss determinations are given, and a theoretical curve that can be used to estimate recharge is derived.

Accuracy of Loss Determinations

Several conditions severely limited the accuracy with which loss from the Little Choconut Creek system could be determined:

(1) Imprecision in streamflow measurement. Most flows larger than 0.5 cubic foot per second (14 liters per second) were measured by pygmy current meter. Most smaller flows were measured by chloride-dilution or volumetric methods; some were occasionally estimated by correlation. The difference between total input and output was less than 8 percent when input exceeded 10 cubic feet per second (283 liters per second). Measurement errors as large as 8 percent are possible with a current meter, but most measurements were made carefully so that errors presumably were less than 8 percent. However, because loss is calculated as a difference between measurements, net error may occasionally be substantial.

(2) Variation in sewer discharge. Two large leaks out of and into municipal sewers that cross the creek were the most serious problems encountered. Rates of leakage varied and could not be measured directly, and the length of time that each leak had existed prior to discovery and repair was not known. Also, industrial wastewater discharged from sewers near Lester Avenue in Johnson City (plate 5) made up a large but highly variable proportion of low flow during working hours; therefore most measurements were made on Sundays, when discharge was smaller and nearly stable.

(3) Time span of measurements and streamflow regimen. Each set of measurements was made during a period of gradually declining natural streamflow several days after the last significant rain. The rate of decline was substantial, especially on sunny summer afternoons, as shown by several paired measurements (table D1) and by stage readings on other dates. Consequently, the relative timing of input and output measurements influenced the determination of loss. For each set of measurements, the times at which the water measured as output at Goudey Station (site J, plate 5) entered the system at Stella or elsewhere were estimated from figure D1, and if necessary the measured inputs were adjusted to those times from the rates in table D1 and available stage records.

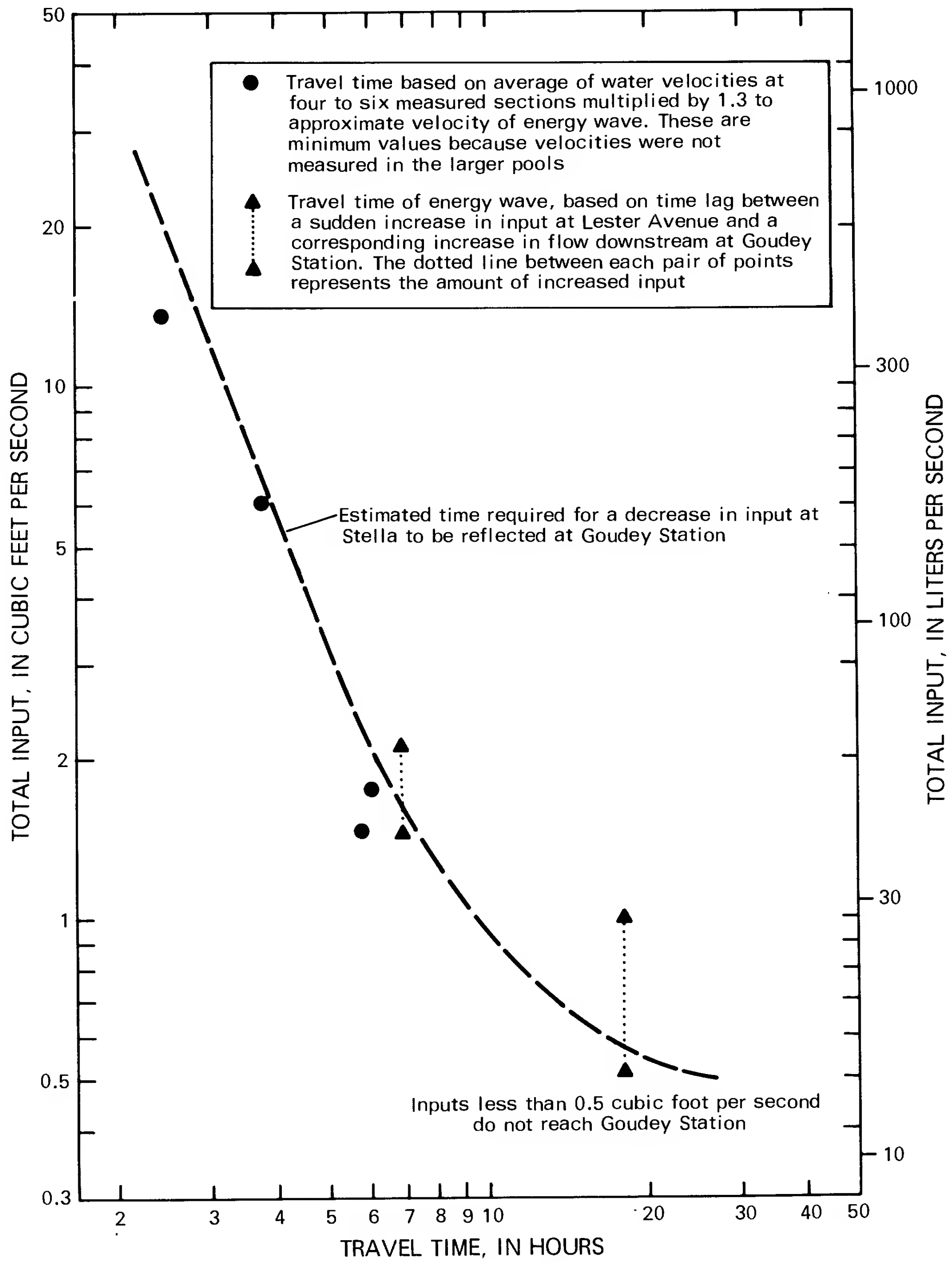


Figure D1.--Travel time, Little Choconut Creek, Stella to Goudey Station. For travel time from Lester Avenue to Goudey Station, multiply by 0.75.

Table D1.--Rate of decline in flow, Little Choconut Creek at Stella[1/]

Date	Period spanned by pair of measurements (hours)	Initial flow (cubic feet per second)	Average rate of decline (percent of initial flow per hour)	Remarks
June 22, 1966	24 hours	1.49	0.7	--
July 14, 1968	24 hours	1.12	.6	--
July 14, 1968	900-1600	1.12	2.8	warm, sunny day
June 6, 1968	1030-1730	9.38	2.9	warm, sunny day
Sept. 16-17, 1968	1800-0800	3.01	1.5	night, steady decline 5 days after large storm
Nov. 1-2, 1968	1630-0845	3.11	2.0	night; at least 4 days after large storm
Sept. 29-30, 1968	1645-0805	.42	No decline	night; average rise of 0.03 percent per hour

1/ Site 1, plate 5.

(4) Evapotranspiration. Measured water loss includes evaporation from the stream surface and transpiration by riparian vegetation. Rough computations based on creek dimensions and data from a lake in North Carolina (Turner, 1966) suggested that evaporation should generally be less than 0.05 cubic foot per second (1.4 liters per second). Transpiration was assumed to be of the same order of magnitude during the growing season, inasmuch as trees were numerous along the channel only between sites H and I (plate 5). If these estimates are approximately correct, evapotranspiration is less than 5 percent of annual water loss as computed below. Accordingly, evapotranspiration is disregarded in this report, and annual water loss from the creek is treated as equivalent to annual recharge to the aquifer.

Recharge is compared with total flow entering the Little Choconut Creek system in figure D2. Despite adjustment for travel time and for miscellaneous errors, the 16 values are not adequate in number, distribution, or reliability to define an input/loss relationship, especially at large inputs. They do suggest, however, that as input increases beyond 2 cubic feet per second (57 liters per second), average loss remains nearly constant.

Factors Controlling Infiltration Rate

The principal factors that could cause time variations in rate of infiltration from Little Choconut Creek are depth to the water table, water temperature, and stream width and depth. In the following paragraphs, each is evaluated.

In this report, depth to the water table is disregarded as an influence on infiltration. Many authors, including Walton and others (1967) and Moore and Jenkins (1966), have pointed out that when the water table declines enough to leave a partially unsaturated zone beneath the streambed, infiltration reaches a maximum rate uninfluenced by water-table fluctuations. During this study, the water table remained 10 to 30 feet (3 to 9 meters) below stream channels in Johnson City, except in small areas near Harry L. Road (plate 8) where the streams begin to cross stratified drift. Water level in a shallow well close to Little Choconut Creek (well 15-55, plate 1) was 4 to 5 feet (about 1.3 meters) below channel grade, and perhaps water levels were also shallow close to Finch Hollow Creek near Harry L. Road. A few measurements suggest that as much as 40 percent of total infiltration from the Little Choconut Creek system occurred in these short reaches. However, we do not know whether ground-water mounds actually intersected the streams, so to simplify subsequent calculations, the streams are assumed to have been totally above the water table in the 1960's. By contrast, records through 1958 show that water levels nearly coincided with stream grade north of the Erie Railroad (plate 7).

As pointed out by Rorabaugh (1956), a decline of 1°C (2°F) in stream temperature will increase the viscosity of the water and lower the infiltration rate by about 2.7 percent. Walton (1963) provides a graph of the relationship between viscosity and temperature. Adjusted infiltration

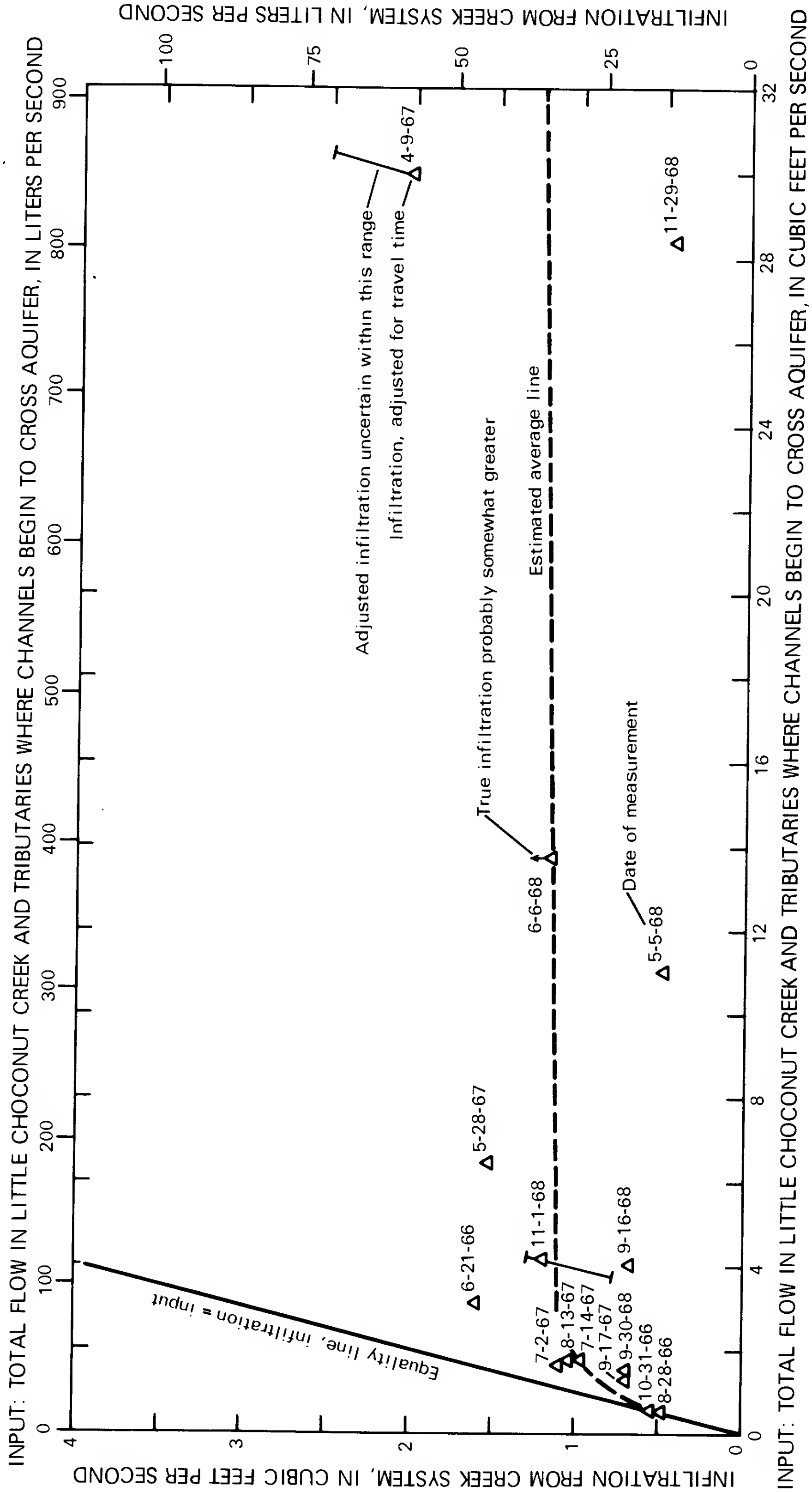


Figure D2.--Infiltration from Little Choconut Creek, adjusted for travel time.

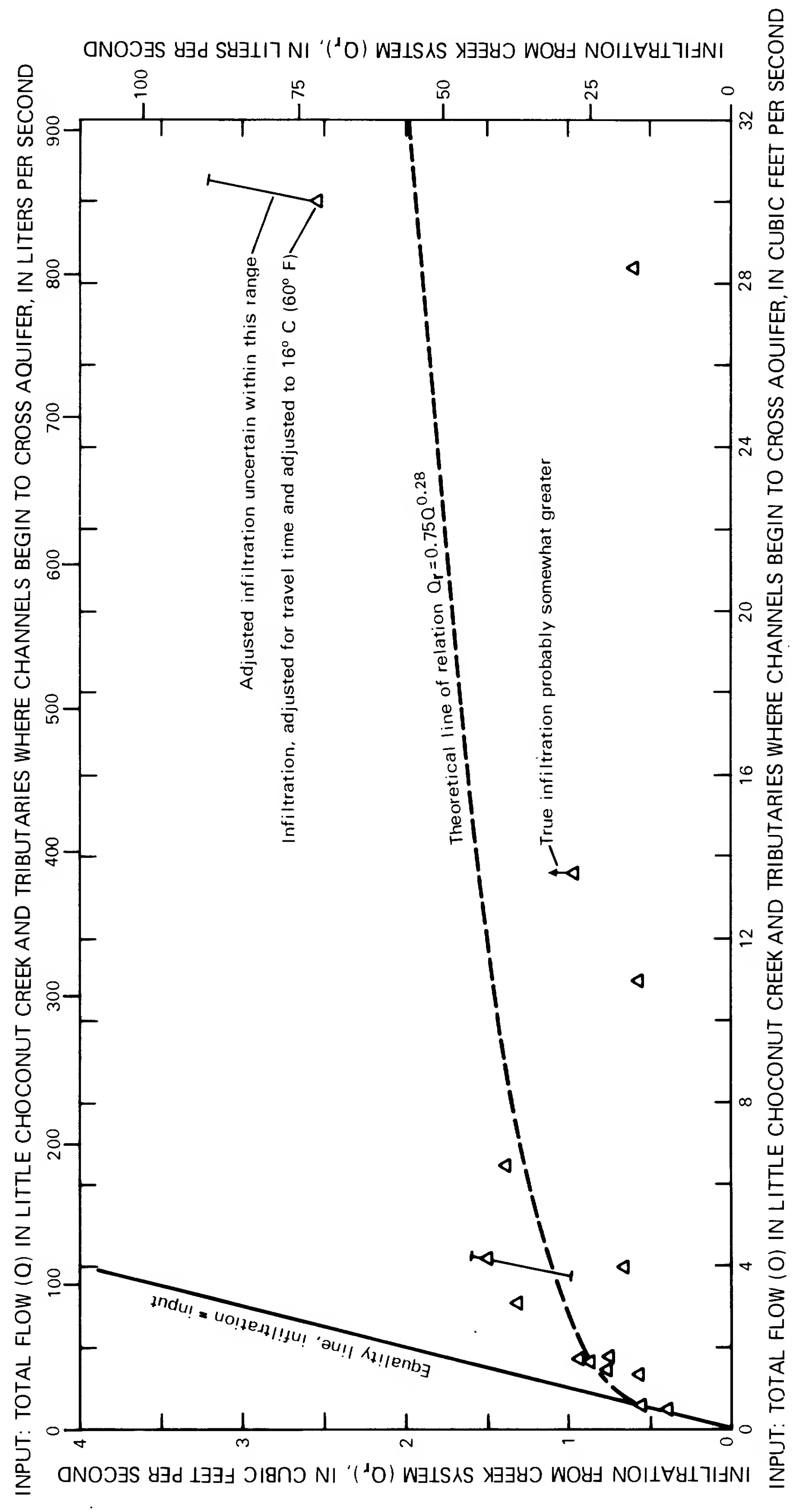


Figure D3.--Infiltration from Little Choconut Creek, adjusted for travel time and temperature.

values from figure D2 were further adjusted from observed temperature to 16°C (60°F) by means of Walton's graph, then plotted against total input in figure D3. Comparing figures D2 and D3, we see that the temperature correction does not improve the scatter of data significantly. However, it does cause the line of relation between loss and input to slope more steeply, because most of the smaller inputs were measured in the summer, when water temperature was above 16°C (60°F).

As streamflow increases, stream width and depth normally increase, thereby providing more streambed area through which infiltration may take place and a greater head to drive the water downward. Bouwer (1965, p. 46) expressed the relationships involved by the following equation:

$$V = \frac{(H_w + L_a - P_c)\ K_a}{L_a} \qquad (1)$$

where V = bulk velocity of infiltration, or specific flux (distance per unit time)
H_w = average water depth
L_a = thickness of layer of low hydraulic conductivity lining streambed
K_a = hydraulic conductivity of that layer
P_c = critical negative pressure at which hydraulic conductivity of underlying soil drops to a low value (see Bouwer, 1964).

Substituting Q_r/Wd for V and rearranging terms, we obtain:

$$\frac{Q_r}{d} = K_a W \ \frac{H_w + L_a - P_c}{L_a} \qquad (2)$$

where Q_r = rate of infiltration (volume per unit time)
W = width of stream
d = length of stream

A similar but less exact equation can be derived from a form of Darcy's law used by Walton (1963) for computing infiltration. Equation 2 indicates that infiltration per unit length of stream is directly proportional to width, and is almost directly proportional to water depth if L_a and $(-P_c)$ are small with respect to depth. According to Bouwer (1964) the value of $(-P_c)$ is close to zero where permeable sand and gravel underlie streambed materials, which is true near Little Choconut Creek. The proper value of L_a is uncertain. Long reaches of Little Choconut Creek are probably underlain by 2 to 5 feet (0.6 to 1.5 meters) of silt (plate 6); one might expect the silt to act as a slowly permeable layer. If so, L_a would be roughly 5 times average H_w, and infiltration would be rather insensitive to changes in water depth. However, the water-loss data do not compel such a conclusion, so perhaps the silt is thinner than estimated above, or infiltration occurs chiefly where the silt is absent.

Leopold and Maddock (1953) and Wolman (1955) have shown that at any particular site, stream depth, width, and velocity are simple power

functions of flow, hence logarithms of depth, width, and velocity increase in constant fractions of any increase in the logarithm of flow. In general, a single set of equations will apply at any site along a particular stream. For example, at any site along Brandywine Creek in Pennsylvania, the following equations (Wolman, 1955) apply very closely:

$$\text{width} = a\ Q^{.04} \tag{3}$$
$$\text{depth} = c\ Q^{.41} \tag{4}$$
$$\text{velocity} = k\ Q^{.55} \tag{5}$$

where a, c, and k are the values of width, depth, and velocity at that site when streamflow (Q) = 1 cubic foot per second (28 liters per second).

During five of the flow measurements in Little Choconut Creek at Goudey Station (site J, plate 5), cross-sectional area perpendicular to flow (Ax) was surveyed or closely estimated in exactly the same place. These data define a line whose equation is

$$A_x = 5.4\ Q^{.28} \tag{6}$$

Similar equations developed for eight small streams north of Owego, New York from data provided by Donald Coates (written commun., 1969) have exponents ranging from 0.19 to 0.37 and averaging 0.27. Their drainage basins are comparable in size and character to Little Choconut Creek. However, their channels are nearly natural at the measurement sites, whereas parts of Little Choconut Creek are lined by dikes and have been relocated, and the alluvium along their channels rests on till and bedrock rather than stratified drift. Pairs of measurements in several places where Little Choconut Creek flows on bedrock at Stella (site 1, plate 5) collectively suggest an exponent of 0.30. Therefore, the exponent of 0.28 developed for site J on Little Choconut Creek (equation 6) appears reasonable.

The coefficient of 5.4 in equation 6 applies only to the site at which it was determined; depth, width, and area coefficients are much smaller for a riffle than for a large pool. Average cross-sectional area of Little Choconut Creek at 1 cubic foot per second (28 liters per second) or any other flow is not known. However, if any poorly permeable materials beneath the channel are thin in relation to water depth, as assumed by Walton (1963), then infiltration per unit length of stream (Q_r/d) would be directly proportional to cross-sectional area (equation 2), or

$$\frac{A_x}{Q_r/d} = \frac{A_{x1}}{Q_{r1}/d} = \text{Constant}$$

where the subscript 1 indicates 1 cubic foot per second (28 liters per second). From equation 6,

$$A_x = A_{x1}\ Q^{.28}$$

Then, by substitution,

$$\frac{A_{x1}\ Q_r/d}{Q_{r1}/d} = A_{x1}\ Q^{.28}$$

$$Q_r/d = (Q_{r1}/d)\ Q^{.28} \tag{7}$$

From figure D3, at an input of 1 cubic foot per second (28 liters per second) infiltration is about 0.75 cubic foot per second (21 liters per second). Stream length (d) may be taken as 1 because all computations pertain to the same reach. Thus, equation 7 becomes

$$Q_r = 0.75\ Q^{.28} \qquad (8)$$

A line representing this equation has been drawn on figure D3 and seems to fit the available data as well as could be expected from the limitations of the data. If surficial silt does indeed act as a slowly permeable streambed layer (L_a) of substantial thickness as compared with depth (H_w), as discussed above, or if ground-water mounds merge with the stream near Harry L. Road, as discussed earlier, then the true line of relation would be somewhat flatter than the line in figure D3.

A subsequent analysis of infiltration from several tributary streams in the Susquehanna River basin where they enter major valleys showed that the rate of infiltration was generally controlled by the hydraulic conductivity of the alluvium or underlying materials rather than by a thin streambed layer of low permeability, and that infiltration was approximately proportional to stream length within the reaches studied but did not correlate with stream width or depth. The reaches studied were generally underlain by gravel alluvium and unaffected by ground-water development, so those empirical conclusions would not necessarily apply to most of the Little Choconut Creek system where channels are incised in silt (or are concrete lined) and the water table is far below the stream as a result of pumping. Nevertheless, they add weight to the suggestion that the true relationship of infiltration to streamflow in Little Choconut Creek may be somewhat flatter than the theoretical curve in figure D3.

Estimated Annual Infiltration

From the theoretical relation of infiltration to flow of Little Choconut Creek (fig. D3) and to water temperature (Walton, 1963, fig. 1), infiltration was computed for the 1967 water year and for the long-term average (table D2). Synthetic flow-duration curves were generated for two partial-record stations and other sites by correlation. The temperature of the Susquehanna River is recorded hourly at Goudey Station near the mouth of Little Choconut Creek, and was similar to temperatures measured in the creek on several dates. Flow records for the nearest long-term continuous-record station representing a small catchment were used to weight monthly mean temperatures at Goudey Station. Final selection of the percent correction for temperature was guided by the fact that infiltration is more nearly proportional to stage than to discharge, as discussed in the section "Factors controlling infiltration rate."

Both Little Choconut Creek and its principal tributary, Finch Hollow Creek, were partly relocated in 1969 to make room for a new highway. Little Choconut Creek now flows through several hundred feet of concrete-lined channel near the former mouth of Finch Hollow Creek. Infiltration from this reach, which was always small because massive silt formed the

Table D2.--Computation of average rate of infiltration from Little Choconut Creek system to Clinton Street-Ballpark aquifer

	Rate of streamflow, or of infiltration, in cubic feet per second: Long-term average (1931-60)	1967 water year
1. Median streamflow (50-percent flow duration) for sources from which flow crosses aquifer [Estimated by correlation (Ku and others, 1975) except for sewers]		
Little Choconut Creek at Stella (U.S. Geol. Survey station 01-5131.90)	3.8	2.5
Finch Hollow Creek at Oakdale (U.S. Geol. Survey station 01-5132.80)	1.1	.7
Minor tributaries	.3	.2
Sewers (estimated from measurements 1966-68)	.5	.5
Total	5.7	3.9

* *

2. Infiltration at 16°C

	Long-term average (1931-60)	1967 water year
From figure D3, corresponding to the total flows estimated above	1.2	1.1

* *

3. Correction of infiltration to actual water temperature

Source of temperature data and method of averaging	Mean temperature, in °C: Long-term average	Mean temperature, in °C: 1967 water year	Percent correction (Walton, 1963, fig. 1): Long-term average	Percent correction (Walton, 1963, fig. 1): 1967 water year
Mean temperature, from hourly measurements, Susquehanna River at Goudey Station, Johnson City.	10.8	11.2	88	88
Discharge-weighted mean temperature, based on mean monthly temperature of Susquehanna River at Johnson City weighted according to 1938-60 and 1967 mean monthly flow of Genegantslet Creek at Smithville Flats.	6.9	7.0	78	78
Stage-weighted mean temperature, calculated as above after converting mean monthly flow to stage	10.0	10.4	86	86
Selected percent correction for temperature	--	--	84	84

* *

4. Estimated average infiltration

	Long-term average (1931-60)	1967 water year
Infiltration at 16°C multiplied by selected percent correction	1.0	.9

channel banks and bottom (plate 6), presumably has been reduced to zero. The lower reach of Finch Hollow Creek was moved eastward to join Little Choconut Creek about 1,000 feet (300 meters) east of its former mouth (plate 5). The new channel is longer and deeper and cuts through the silt to the underlying gravel in places, hence larger infiltration rates can be expected. Three flood-control dams upstream were completed in 1969, and a fourth was planned; they will alter flow distribution and should reduce silt load in water crossing the aquifer. Other changes in flow or channel conditions may be expected in this urban area. Hence, periodic revaluation of infiltration may be warranted.

* * *

APPENDIX E: RECHARGE FROM PRECIPITATION

Method of Computation

Computations of recharge from precipitation for three different periods ending on October 6, 1967 are summarized in table E1. The principal steps in computation were as follows:

1) Water-level contours and flow lines were drawn for two dates (plates 7, 8). From these maps and other measurements, contours of net change in water level were drawn for each computation period.

2) Net change in storage was computed for each cone of depression (or segment thereof) that did not intersect the Chenango or Susquehanna rivers.

3) Infiltration from Little Choconut Creek was calculated from table D2 and was subdivided roughly in proportion to stream length within each cone of depression.

4) Pumpage from each cone of depression was copied from well-owners' records, or (for a few small wells) estimated from pump capacity and the owner's recollection of operating time.

5) Net change in storage, infiltration, and pumpage were combined to obtain recharge per square mile from precipitation for each cone of depression.

Table E1.--Computation of recharge from precipitation

[Columns 3-6 are in million gallons]

1	2	3	4	5	6	7	8	9	10		
									Recharge per unit area (col. 8 ÷ col. 9)		
Time period	Principal wells 1/	Water pumped from aquifer 2/	Net change in storage 3/	Infiltration from Little Choconut Creek	Recharge from precipitation (col.3 ± col.4 -col.5)	Days in time period	Mean daily recharge (col.6÷col.7) (million gallons per day)	Area of stratified drift 4/ (square miles)	million gallons per day per sq. mile	inches per year	inches in time period
Oct. 18, 1966 to Oct. 6, 1967	GAF 3, 5	359	+56	--	415	353	1.18	1.18	1.00	21	20
	Johnson City Ballpark	714	+84	122	676	do.	1.92	1.29	1.49	31	30
	Johnson City 5	125	+14	45	94	do.	.266	.21	1.27	27	26
	(subtotal) 5/	(1,198)	(+154)	(167)	(1,185)	do.	(3.36)	(2.68)	(1.25)	(26.5)	(25.5)
	Johnson City 1, 2, 3	516	+28	43	501	do.	1.42	.92	1.54	32	31
Apr. 25, 1967 to Oct. 6, 1967	GAF 3, 5	294	-180	--	114	164	.695	1.18	.59	12.4	5.6
	Johnson City Ballpark	316	-53	71	192	do.	1.17	1.29	.907	19	8.5
	Johnson City 5	91	-33	23	35	do.	.213	.21	1.01	21	9.4
	(subtotal) 5/	(701)	(-266)	(94)	(341)	do.	(2.08)	(2.68)	(.776)	(16)	(7.2)
	Johnson City 1, 2, 3	244	-123	20	101	do.	.616	.92	.670	14	6.3
Sept. 25, 1958 to Oct. 6, 1967	GAF 3, 5	3,384	-197	--	3,187	3,296	.967	1.18	.82	17.2	155
	All wells 6/	16,300	-1,070	2,120	13,100	do.	3.97	3.60	1.11	23.3	210

1/ Includes minor withdrawals from several wells; see appendix B.

2/ 37 to 52 percent of total withdrawal from Johnson City wells 1,2,3, corresponding to portion of flow net not intersecting Susquehanna River; see appendix B. Total withdrawal for other wells listed.

3/ Gravity yield taken as 0.16 for period April to October 1967, 0.2 for other periods.

4/ For Johnson City wells 1, 2, and 3: area of 3/8 of 1967 flow net not intersecting Susquehanna River (pl. 8). For other wells, area of entire cone of depression. Includes thinly saturated stratified drift north and south of aquifer but draining to it.

5/ Includes all wells listed except Johnson City 1, 2, 3.

6/ Includes Johnson City 1, 2, 3 as well as other wells listed.

6) Results were compared, and the shape of the flow net was adjusted where possible to improve consistency. Although plate 8 indicates that in October 1967 about 37 percent of the pumpage from Johnson City wells 1 to 3 was derived from the aquifer rather than from or beyond the Susquehanna River, the position of the flow lines that define this percentage could easily be shifted if transmissivity were assumed to be nonuniform, so this well field was not considered in computing recharge per square mile for 1966-67.

Accuracy of Results

The computation for October 1966 to October 1967 is the most reliable. Water-level measurements were more nearly simultaneous in October 1966 than at the start of the other two periods, pumpage records are more complete, and net change in water level was least.

Recharge from precipitation from April 25 to October 6, 1967, a 5.3-month period approximately coinciding with the growing season, was computed to be 27 percent of the annual total. Recharge is often assumed to be slight during the growing season. However, substantial recharge did occur in May 1967, as shown by a rise in ground-water levels (fig. 6), reflecting 4.2 inches (11 centimeters) of rain early in the month. Also, some rain in March or April may not have reached the water table by April 25, and a small lateral flow from bedrock and till continues during the summer.

For each time period, recharge per square mile of stratified drift computed for GAF wells 3 and 5 in Binghamton is roughly 75 percent of that for wells in Johnson City. There are several possible explanations for this inconsistency.

1) Perhaps it is the net result of small errors in measurement or estimate of pumpage, drawing flow nets, estimating gravity yield, proportioning infiltration from Little Choconut Creek among various wells, etc. Note that pumpage from the Ballpark well was not metered (appendix B).

2) Perhaps the area of stratified drift assigned to GAF wells 3 and 5 is too large, hence the unit recharge rate too small. Half of that area lies outside the aquifer (whereas only small areas outside the aquifer are assigned to other wells) and locally its southern boundary (plate 6) is not precisely known. Sand and gravel in these marginal areas is only thinly saturated, and water-level fluctuations do not reflect those in the aquifer.

3) Perhaps infiltration from Little Choconut Creek is underestimated. Data supporting the curve in figure D3 used to predict infiltration are admittedly weak and scattered. However, to equalize estimated recharge from precipitation in Johnson City and Binghamton solely by increasing the estimate of infiltration from the creek in Johnson City would require more than twice as much infiltration, and neither the curve in figure D3 nor the calculated 50-percent flow duration for 1967 (table D2) could be raised this much without greatly exceeding the data.

4) Perhaps recharge from precipitation is genuinely less in Binghamton than in Johnson City because of Binghamton's more extensive urban development.

Frequency Distribution of Annual Recharge

To determine directly the variation from year to year in recharge from precipitation would require many annual sets of water-level measurements in numerous wells. However, the variation in annual precipitation and runoff are known and may be used to estimate the variation in annual recharge.

Figure E1 shows the probability distribution of annual runoff from Genegantslet Creek and of precipitation 12 miles (19 kilometers) to the east from 1939 through 1967. Comparison of the two trend lines suggests that the difference between rainfall and runoff is nearly the same in both wet and dry years. Although long-term mean rainfall and runoff both vary more than 10 inches (25 centimeters) from east to west across the Susquehanna River basin, the difference between them appears to vary less than 2 inches (5 centimeters) across the basin (Ku and others, 1975). The difference between precipitation and runoff in a given locality over a year or more largely represents loss by evapotranspiration. In the area of the Clinton Street-Ballpark aquifer, nearly all precipitation becomes either evapotranspiration or ground-water runoff. Accordingly, net recharge to the aquifer from precipitation should vary over the years by about the same amount as the variation in local precipitation.

The probability distribution of recharge from precipitation to the Clinton Street-Ballpark aquifer is inferred in figure E2. Recharge calculated for 1967 and 1958-67 (table E1) was plotted at the same percent of time as annual rainfall at Binghamton for those years. Two probability distribution lines were drawn, both of which assume constant evapotranspiration loss. One line, based on GAF wells 3 and 5 in Binghamton, suggests a median annual recharge of 22 inches (56 centimeters); the other, based on several wells, suggests a median annual recharge of 27 inches (69 centimeters). Both values exceed the long-term mean runoff of about 19 inches (48 centimeters) from this part of the Susquehanna basin (Ku and others, 1975) probably because recharge from precipitation as calculated in this report includes a small amount of lateral ground-water inflow from adjacent till and bedrock, and because transpiration from the area studied must be less than the basinwide average. Urban development limits plant density. Furthermore, the water table was 15 to 60 feet (4.5 to 18 meters) below land surface nearly everywhere in this aquifer throughout the period of study, deeper than the roots of many plants, whereas throughout the basin, many valley areas have a water table only 5 to 15 feet (1.5 to 4.5 meters) below land surface, and in the upland areas that make up 85 percent of the basin, most runoff flows toward streams through the top foot or two of glacial till. Ground-water evapotranspiration has been calculated to be 20 and 22 percent of total evapotranspiration in glaciated basins of Pennsylvania and Connecticut (Olmsted and Hely, 1962; Randall and others, 1966), but was probably much less in the Clinton Street-Ballpark aquifer during the period of study.

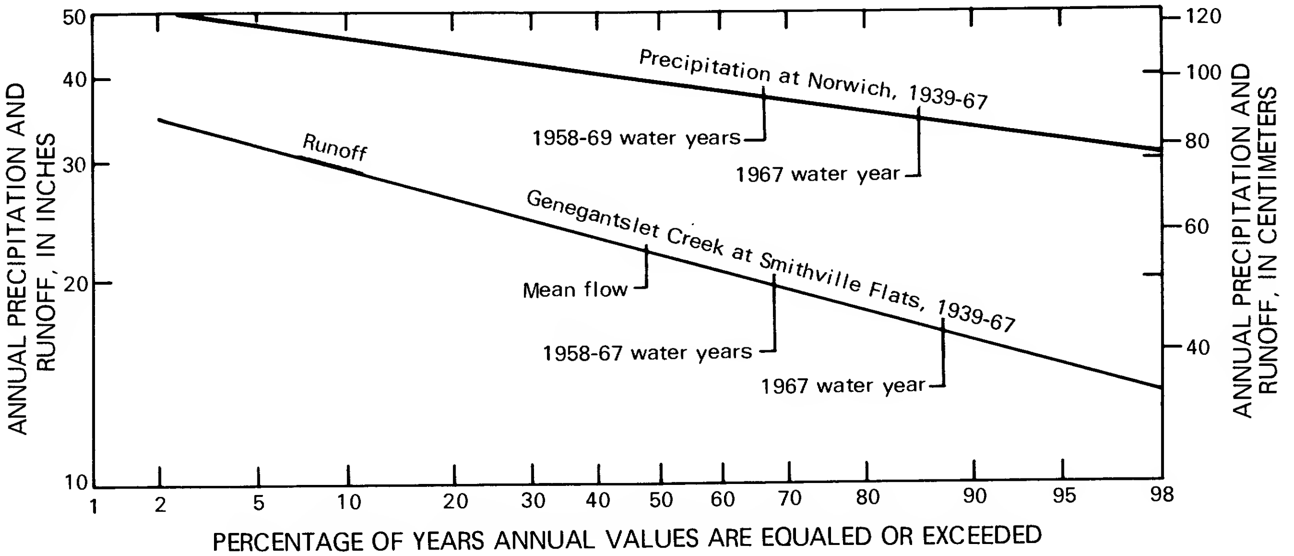


Figure E1.--Probability distribution of annual precipitation at Norwich and runoff in Genegantslet Creek.

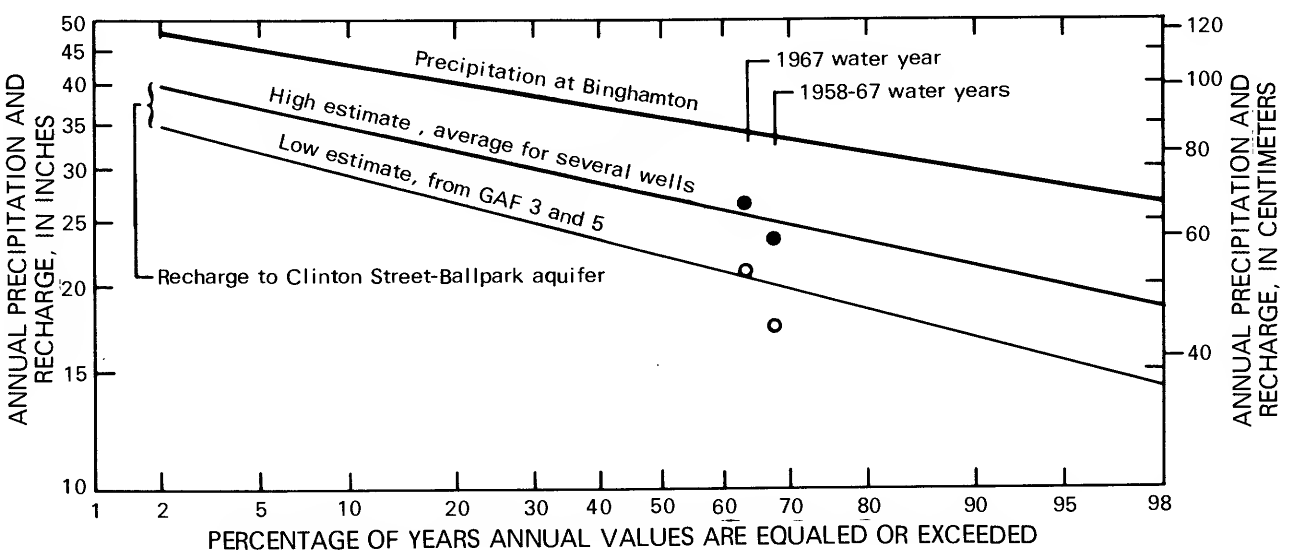


Figure E2.--Probability distribution of annual precipitation at Binghamton and of recharge to the Clinton Street-Ballpark aquifer.

Effect of the Urban Environment

Assuming that lines in figure E2 provide "high" and "low" estimates of median annual recharge from precipitation, the water budgets in table E2 were constructed to indicate the sources and disposal of recharge to the Clinton Street-Ballpark aquifer exclusive of induced recharge from the major rivers. Because impervious pavement and buildings covered a substantial part of the aquifer, ground-water recharge from precipitation might be expected to be substantially less in 1967 than under natural conditions. However, table E2 suggests that the degree of urbanization and ground-water development prevailing in Binghamton and Johnson City in 1967 caused a greater reduction in evapotranspiration than in recharge from precipitation, hence at least as much water was available for use under developed conditions as natural conditions. The 21 to 37 percent reduction in evapotranspiration suggested by table E2 may be too great, but the figures on which it is based are reasonable estimates consistent with available data. The rate of leakage from water mains or sanitary sewers is unknown and is not separately mentioned in table E2; perhaps recharge from this source was significant in 1967 and compensated for a modest reduction in natural recharge from precipitation.

Comparison with Other Areas

Ground-water recharge has been computed from water-budget studies for several areas in the glaciated northeastern United States. Most of these studies (for example, Meinzer and Stearns, 1929; Parker and others, 1964; Randall, 1964) present average recharge rates for entire basins above one or more gaging stations; such averages obscure wide differences in recharge per square mile between small areas of permeable stratified drift and much larger areas of poorly permeable till and bedrock, hence are of little practical value.

Recharge in a partially urbanized area of central Long Island, New York that is underlain largely by stratified drift was calculated by Cohen and others (1968) to average 23 inches ± 6 inches (58 ± 15 centimeters) and was derived entirely from precipitation. Calculations were based on measured precipitation minus evapotranspiration from land surface and unsaturated soil (computed by a modified Thornthwaite method) minus storm runoff.

Recharge from precipitation to stratified drift in eastern Connecticut was calculated to average 21 inches (53 centimeters) by Randall and others (1966) on the basis of streamflow hydrograph separation at several gaging stations. This value does not include lateral ground-water flow from adjacent till and bedrock (computed separately for any specific area), nor does it include recharge that became ground-water evapotranspiration. If rough estimates for these quantities are added, recharge to stratified drift exclusive of infiltration from streams averages about 28 inches (710 millimeters), and bears about the same relation to rainfall and runoff as does the low estimate for the Clinton Street-Ballpark aquifer.

* * *

Appendix E

Table E2.--Water budgets, exclusive of induced recharge from rivers.
[Budgets are computed for average conditions: input must equal output, no net change in storage]

		Budget item	Annual amount, expressed to nearest inch, averaged over area of aquifer plus marginal sand and gravel 1/		
			Natural conditions	Urban conditions as of 1967	
				Low estimate of recharge	High estimate of recharge
Total water budget	Input	Precipitation on stratified drift	2/ 37	2/ 37	2/ 37
Total water budget	Input	Lateral inflow from till and bedrock	3/ 3	3/ 1	3/ 3
Total water budget	Input	Infiltration from Little Choconut Creek	4/ 2	5/ 3	5/ 3
Total water budget	Output	Storm runoff from impervious and water-saturated surfaces	6/ 0	6/ 2	6/ 2
Total water budget	Output	Total evapotranspiration	7/ 19	8/ 15	8/ 12
Total water budget	Output	Ground-water runoff and (or) pumpage	8/ 23	9/ 24	9/ 29
Ground-water budget	Recharge (input)	From Little Choconut Creek	4/ 2	5/ 3	5/ 3
Ground-water budget	Recharge (input)	From precipitation, including lateral inflow and leaky pipes or sewers	9/ 25	10/ 22	10/ 27
Ground-water budget	Discharge (output)	Runoff	8/ 23	--	--
Ground-water budget	Discharge (output)	Pumpage	--	9/ 24	9/ 29
Ground-water budget	Discharge (output)	Evapotranspiration of ground water	11/ 4	4/ 1	4/ 1

Footnotes to Table E2

1/ Area of aquifer is 2.99 square miles (7.75 square kilometers); total area including marginal sand and gravel is 3.94 square miles (10.20 square kilometers).

2/ U.S. Weather Bureau (1965).

3/ Most lateral inflow from bedrock probably originates within a mile north of the aquifer (inferred from topography). Recharge to bedrock through silty till may be about 0.4 inch (1 centimeter) annually (Heath, 1964), in which case lateral inflow from bedrock is no more than 0.4 inch expressed over the aquifer. Also, however, for each lineal mile (1.6 kilometer) of valley there is about 0.3 square mile (7.8 square kilometer) of till-covered hillside from which storm runoff drains toward the aquifer; although its magnitude may approach regional mean runoff of 19 inches (48 centimeters) or nearly 6 inches (15 centimeters) expressed over the aquifer, much of it probably reaches streams rather than infiltrating to the aquifer through gravel at the base of the hillside, especially in 1967 when streets with storm sewers crossed most of the hillsides.

4/ Estimated.

5/ From table D2; 1 cubic foot per second (28 liters per second) distributed over 3.94 square miles (10.2 square kilometers) is about 3.4 inches (8.6 centimeters). Infiltration should have been less under natural conditions when the water table was much higher.

6/ Storm runoff from stratified drift on Long Island ranges from about 0.5 inch (1.3 centimeters) under near-natural conditions to about 2 inches (5 centimeters) under relatively urban conditions (Pluhowski and Kantrowitz, 1964, p. 30-35). Although rainfall is greater on Long Island, the most urbanized basins studied by Pluhowski and Kantrowitz appear from the topographic map to have had more open space than the Clinton Street-Ballpark aquifer at the time of study, and the Long Island runoff values do not include storm runoff diverted to recharge basins.

7/ From regional maps by Ku and others (1975).

8/ Calculated, to balance other items of input and output in total water budget.

9/ Calculated, to balance recharge and discharge in ground-water budget.

10/ From figure E2.

11/ 20-22 percent of total evapotranspiration, according to Olmstead and Hely (1962), Randall and others (1966).

APPENDIX F: RECHARGE INDUCED FROM CHENANGO AND SUSQUEHANNA RIVERS

East End of the Aquifer

Detailed lithologic logs, water levels, temperature, and chemical analyses from wells in a small area at the east end of the Clinton Street-Ballpark aquifer all suggest an aquifer geometry such as that shown by cross sections (fig. 3) and a block diagram (fig. F1). At least 38 feet (12 meters) of bright sand and gravel are generally present at the top of the aquifer. In a narrow east-west belt, part of this bright sand and gravel is warped downward beneath 50 feet (15 meters) of silt and makes imperfect contact with deeper drab sand and gravel. Induced recharge from the Chenango River moves readily through the shallow gravel and down the north limb of the downwarped bright-gravel unit toward GAF well 7 (fig. F1). Induced infiltration also moves toward GAF well 7 from the southeast, but so slowly that seasonal extremes of water temperature 200 to 400 feet (60 to 120 meters) southeast of this well lag 6 to 8 months behind corresponding extremes of river temperature. The deep drab sand and gravel northeast of GAF 7 (fig. 3) has not shown temperature changes and thus may be poorly connected to the river. The amounts of water reaching GAF 7 via these different flow paths may be estimated from a striking contrast in temperature observed in GAF 7 on September 4, 1967, after the pump had been shut down for 27 hours (fig. F2). Temperatures at depths less than 100 feet (30 meters) were 17.5° to 19.3°C (63.6° to 66.8°F), similar to those in the shallow gravel at nearby observation wells on this date. At depths of 104 to 107 feet (31.7 to 32.6 meters), temperatures averaged 10.3°C (50.6°F), slightly cooler than in the deep gravel nearby. Water pumped was 17.2°C (63.0°F) just before shutdown. By proportion, it is estimated that 78 percent of the water pumped must have come from the shallow sand and gravel via the downwarped unit immediately northeast of GAF 7. Most of the remaining 22 percent probably infiltrated through the eastern part of the river bed, moved directly downward, and approached GAF 7 from the southeast through the deep gravel (fig. F1), inasmuch as well 27-47 (plate 1) is the only one of several nearby observation wells at which temperatures had been as low as 10.3°C (50.6°F) in the deep gravel during the previous few months, which were the coldest in an annual temperature cycle at that depth.

Over Labor Day weekend 1967, all GAF production wells were shut down for at least several hours. Water levels measured during this period in wells near the east end of the aquifer may be classified in three groups:

1) shallow wells, and deep well 27-47: water levels 0.8 foot (0.24 meter) below river level with GAF well 7 idle, 1.3 feet (0.40 meter) or more below river level with GAF well 7 pumped.

2) GAF well 7 and deep observation wells nearby: head 6 ± 2 feet (1.8 ± 0.6 meters) below river level with GAF well 7 idle.

3) GAF well 8 and 9, and a nearby observation well: head roughly 30 feet (9 meters) below river level with all pumps idle.

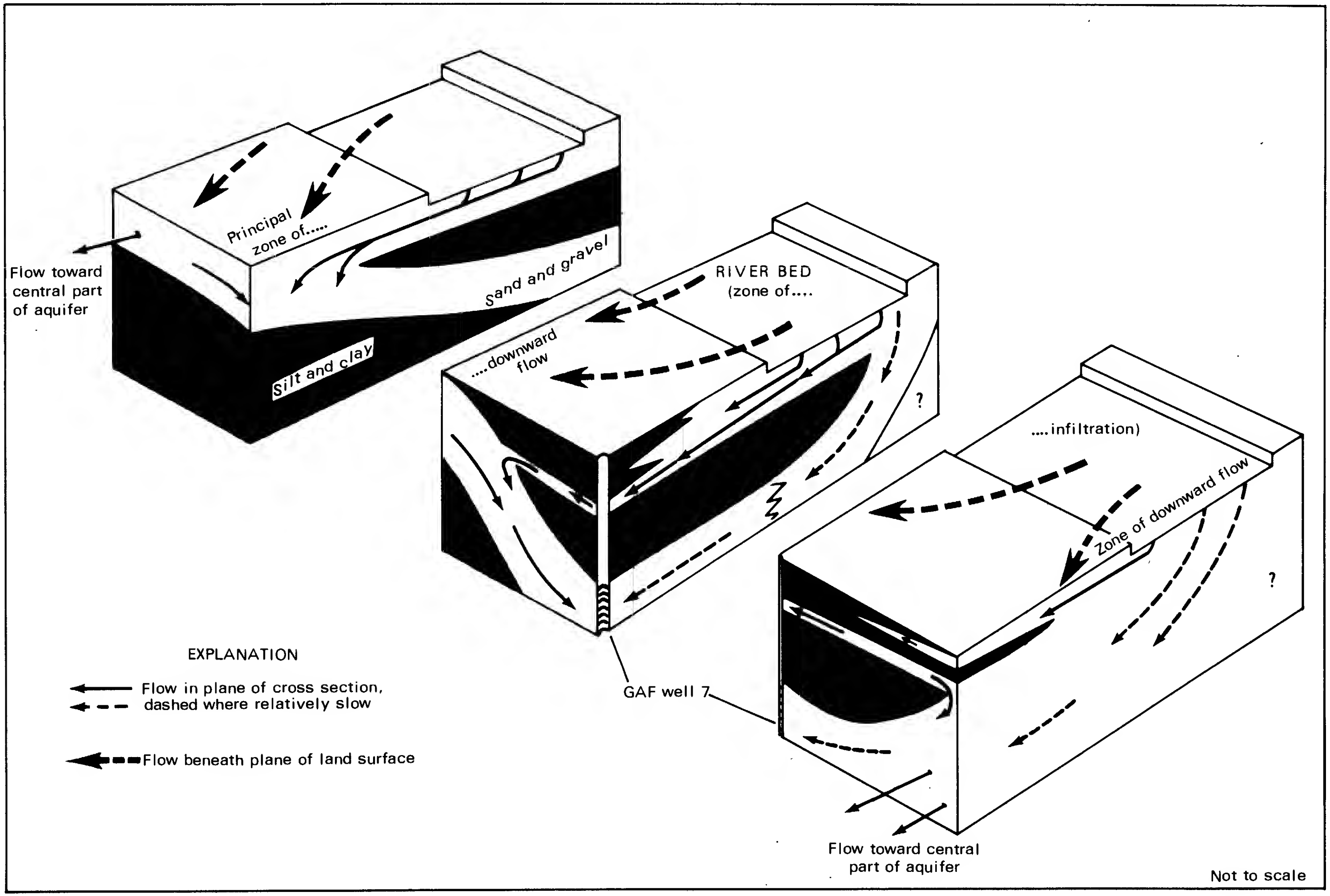


Figure F1.--Movement of induced recharge near GAF well 7, east end of aquifer. The 3 blocks represent adjacent segments of the earth, sliced and pulled apart so that the cross sections between them could be shown.

Recovery following shutdown of GAF well 7 was virtually complete within 16 hours; thereafter, water levels in the first two groups of wells declined nearly parallel to a decline in river stage. Under these circumstances, water levels (fig. F3) reflect movement of induced recharge from the Chenango River west into the interior of the Clinton Street-Ballpark aquifer to replace storage depleted by previous pumping. The distribution of water levels suggests that the head (or potential energy) lost in moving water from the river into the shallow gravel layer is only one-seventh as great as that lost in moving water from the shallow gravel down into the basal gravel, which in turn is only one-fourth of the loss in moving water westward to the vicinity of GAF wells 8 and 9. Therefore,

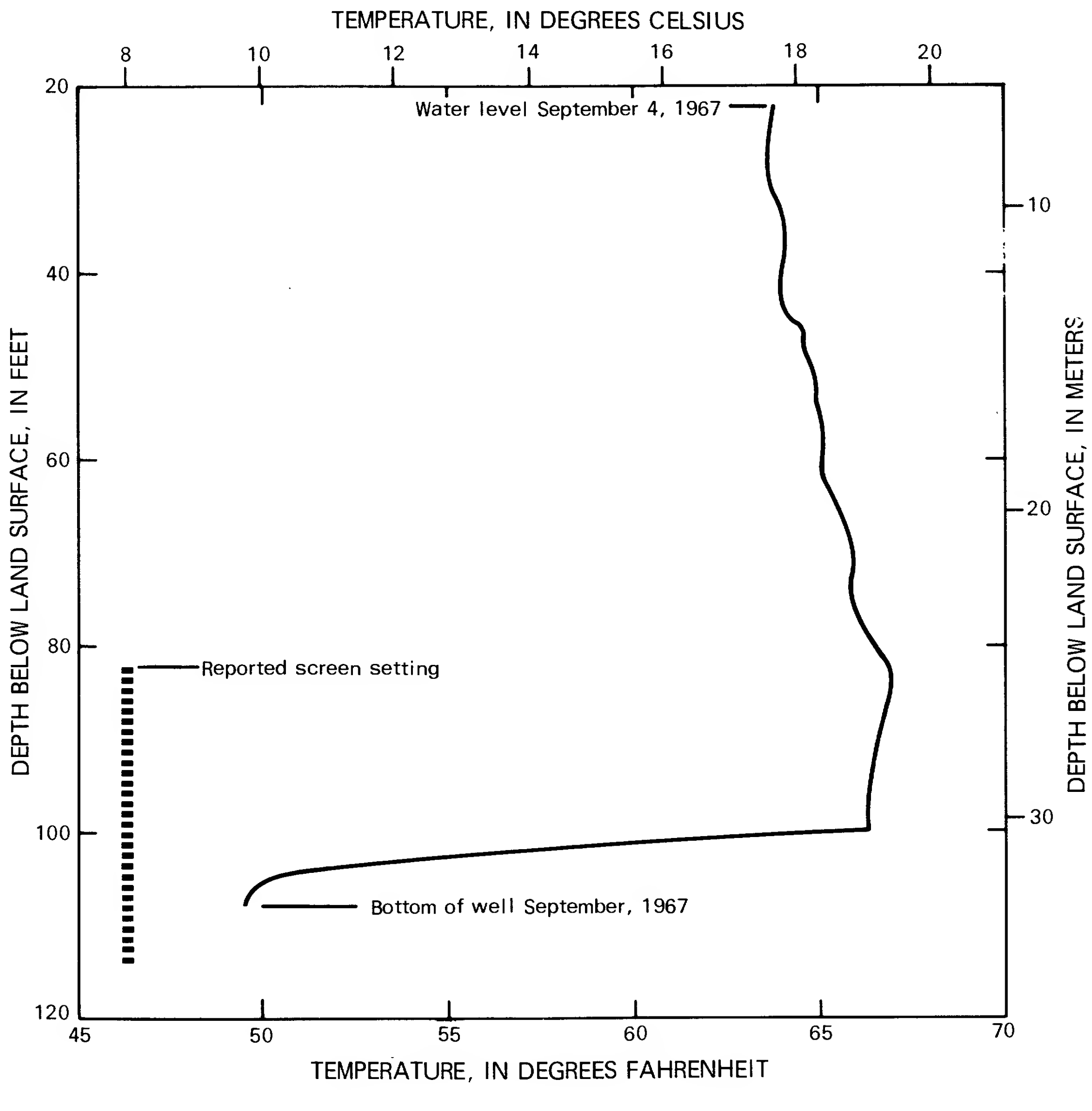


Figure F2.--Temperature profile in GAF well 7, September 4, 1967.

if a battery of screened wells were installed east of Front Street, they should be able to lower the head in the shallow gravel layer much more effectively than the existing wells do, and would thereby cause several times as much river water to infiltrate. River depth at low flow in September 1967 was 4 to 8 feet (1.2 to 2.4 meters) in the central and western parts of the channel, 2 to 4 feet (0.6 to 1.2 meters) in the eastern part. Drawdown beneath the channel is unknown, but probably did not exceed 1.2 feet (0.37 meter) near the west bank and 0.6 foot (0.18 meter) near the east bank, judging from the measurements quoted above in shallow wells west of the river. If so, induced recharge could be increased to at least five times its September 1967 value by placing wells east of Front Street and pumping enough to lower head in the aquifer beneath the stream to a level below the streambed.

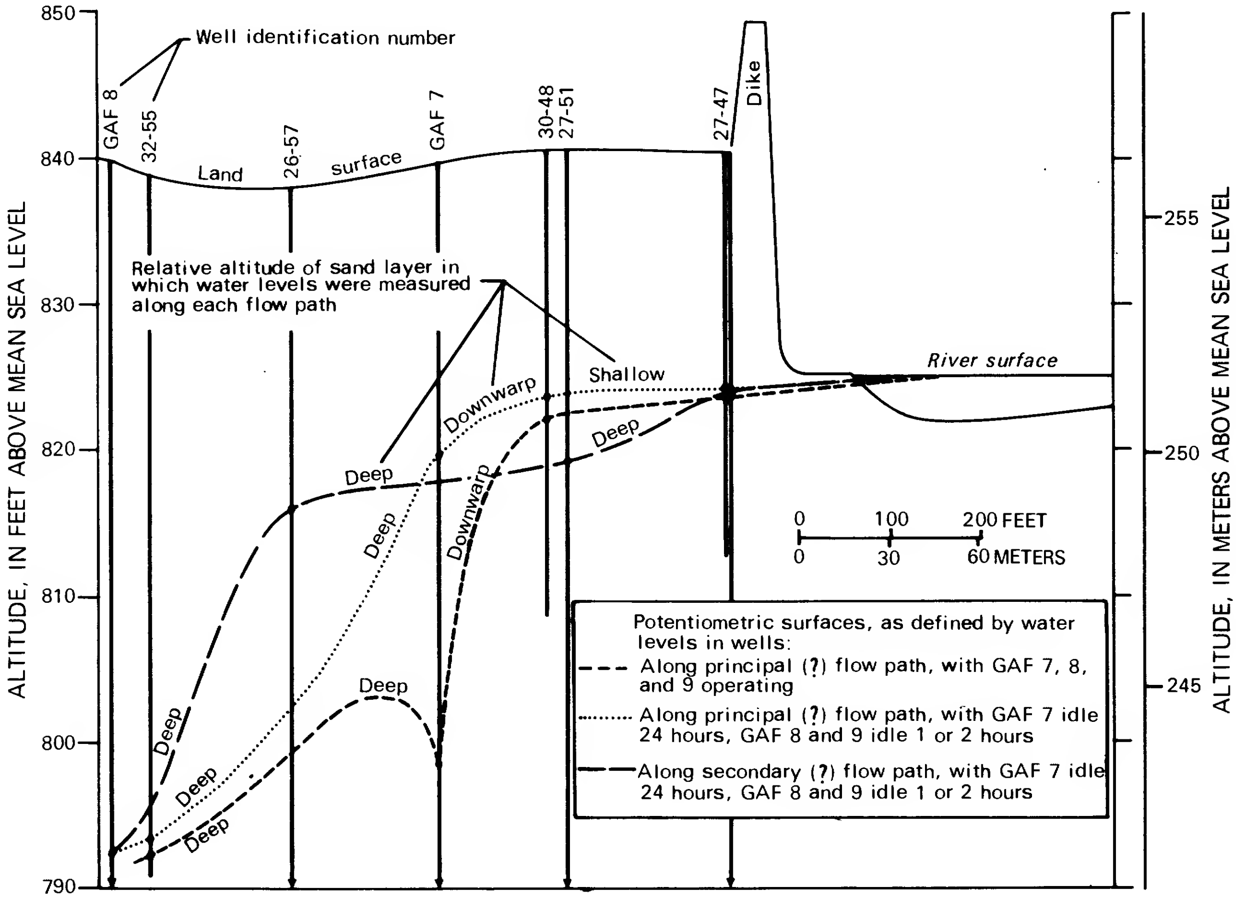


Figure F3.--Water-level profiles at east end of aquifer, September 3-4, 1967. Water levels measured in selected observation wells are projected to a line of section drawn between wells 27-47 and GAF well 8. For location of wells, see plate 1. For geometry of the downwarped sand layers that constitute the inferred flow paths, see fig. F1 or fig. 3 B-B' and B-C.

Induced recharge at the east end of the Clinton Street-Ballpark aquifer averaged about 3.3 million gallons per day (12.5 million liters per day) in 1966-67. This figure was computed by taking pumpage from the area that includes GAF wells 6-10 (appendix B), then subtracting recharge from precipitation (estimated per square mile from GAF wells 3 and 5, table E1) and net change in storage. It includes an unknown but probably small proportion of ground-water underflow from beyond the river. Average width of the Chenango River is approximately 350 feet (107 meters) in a reach 4,500 feet (1,600 meters) long at the end of the aquifer. Thus, streambed area is about 1.6 million square feet (150,000 square meters), and because not quite all the 3.3 million gallons per day (12.5 million liters per day) came from the river, the average infiltration rate in 1966-67 was no more than 2.1 gallons per day per square foot (85 liters per day per square meter). A fivefold increase, to 10.5 gallons per day per square foot (430 liters per day per square meter), as suggested in the previous paragraph, would still be below the few published rates of streambed infiltration measured where pumping had lowered water level below the streambed (Moore and Jenkins, 1966). The streambed here has been excavated and relocated several times (fig. 8), which should make it more permeable than an undisturbed natural streambed (Randall, 1970). As of September 1967, the bed was chiefly gravel with a slight surface accumulation of very loose silt, except close to the banks where as much as 2 feet (0.6 meter) of organic silt covered the bottom. Therefore, it seems likely that a fivefold increase in induced recharge to 16.5 million gallons per day (62 million liters per day, including underflow) is feasible at the east end of the aquifer.

Computations using a simple mathematical model (Walton, 1962, p. 24, 55) suggest that it might be possible to obtain as much as 16.5 million gallons per day (62 million liters per day) from a line of wells 200 feet (60 meters) from the river bank, 50 feet (15 meters) in depth in the shallow aquifer (fig. F3) and deeper to the north and south where the aquifer is not interrupted by thick silt-clay lenses and has higher transmissivity (fig. F3, plate 3). However, this aspect should receive careful reevaluation when new wells are designed and installed.

West End of the Aquifer

Induced infiltration at the west end of the aquifer probably averaged about 2.2 million gallons per day (8.3 million liters per day) in 1966-67. This figure does not include ground-water underflow from beyond the river because it was computed by taking pumpage from segments of the flow net that intersect the river, then subtracting recharge from precipitation and net increase in storage within those entire segments, including the parts beyond the river (plate 8). The "high estimate" of recharge from precipitation of 26.5 inches (67 centimeters) per year in 1966-67, based on average data for several wells (table E1) was used. The area of river bottom influenced by Johnson City wells 1-3 (plate 8) was about 2.2 million square feet (205,000 square meters), hence an infiltration rate of 1.0 gallon per day per square foot (40 liters per day per square meter) is indicated. This rate is substantially smaller than that computed for the Chenango River. Even if the lowest 1966-67 estimate of recharge

from precipitation (based on GAF wells 3 and 5, table E1) were applied over the entire flow net, the infiltration rate here would be only 1.3 gallons per day per square foot (53 liters per day per square meter). There are two independent reasons for expecting a smaller infiltration rate here than along the Chenango River:
(1) Temperature records from Johnson City wells 1-3, although imperfect, suggest a smaller annual temperature variation than that of GAF well 7, despite heavier withdrawals closer to the river bank.
(2) Riverbed conditions. The river bed here is chiefly gravel, but its natural stratification has not been disturbed by excavation as much as the bed of the Chenango River near Front Street. Also, test borings near the center and left bank 700 feet (210 meters) south of Johnson City well 2 penetrated fine sediment from 0 to 40 feet (12 meters) that would not yield water, and there is some evidence that silt layers along the right bank limit infiltration. (See section "Sanitary quality.")

Infiltration may have already reached its maximum rate over a large part of the streambed influenced by Johnson City wells 1-3. River depth at low flow in September 1967 averaged 4 or 5 feet (about 1.4 meters), and the contours in plate 8 suggest an average drawdown of about 8 feet (2.4 meters) beneath the channel. However, contours near the river are controlled only by measurements made by air line in Johnson City wells 1-3, which are screened between 30 and 70 feet (9 and 21 meters) below river level in an area of large withdrawals. Therefore, the contours may be inaccurately drawn near the river and (or) may reflect head deep within the aquifer rather than immediately below the streambed, in which case increased pumpage would result in a steeper gradient and thus increased infiltration through the streambed and underlying sediment. A maximum infiltration rate of only 1 or 2 gallons per day per square foot (40 or 80 liters per day per square meter) seems small if aquifer transmissivity ranges from 10,000 to more than 50,000 feet squared per day (900 to more than 4,600 meters squared per day, plate 3). However, until better definition of water levels and flow patterns near and beneath the streambed are obtained, it seems unwise to predict a large potential for increased infiltration. All things considered, an additional 2 million gallons per day (7.6 million liters per day) could probably be obtained by induced recharge from a new well or wells 2,000 feet (600 meters) northwest of Johnson City wells 1-3 near the river. With better data, it might be possible to raise this estimate.

Alternative Methods of Estimating Induced Recharge

The primary method used in this report to estimate the volume of induced recharge reaching each well is to compare estimated recharge from other sources with total pumpage, adjusted for storage changes. A computation based on water temperature in GAF well 7 was described earlier. If a sample of the water pumped is analyzed chemically and is assumed to be a mixture of average ground water and average river water, each with chemical quality as described in table 3, the volume derived from each source may be calculated by proportion. Results by these three methods are compared in the following table:

Appendix F

Well (plate 1)	Percentage of river infiltration in water pumped, estimated from			
	Water levels and pumpage	Temperature of water	Hardness of water	Chloride concentration
GAF well 7	>98	>89	77	69
Johnson City Wells 1, 2, 3	57	--	49	68

The chemical method is less precise because the average values in table 3 are only approximations and because possible increases in mineral content as river water passes through the aquifer are ignored, as are seasonal changes in quality of river water, which may coincide with variations in infiltration rate. However, the collective results of all three methods indicate that the proportion of induced infiltration is substantial, especially at the east end of the aquifer.

* * *

Table G1.--Wells in which water-level measurement could be made as of 1971

[Wells listed from east to west within successive 1-minute strips of latitude. Depths, logs, and other specifications of these wells are given in Randall (1972).]

Well identification and location[1/]				Measuring point				
Owner's well number	Latitude	Longitude	Owner[2/]	Description	Elevation above(+) or below(-) land surface[3/] (feet)	Altitude (feet)	Source of altitude measurement[4/]	Description of well location; remarks
	4206 34	7554 42	Cutler Ice Co.	Top of casing, in pit	-2.6	840	TM	Pumphouse alongside Cutler Ice Co. building. Pumped continuously.
	15	45	U.S. Geol. Survey	Top 2-inch hole in plug atop 6-inch casing	0	845	TM	7 feet north of base of railroad track support
	31	46	do.	Top of 6-inch coupling	+ .9	840.5	USGS	46 feet from toe of dyke, behind 302 Front Street
	27	47	do.	Top threads on 6-inch coupling	+1.8	843.5	do.	Toe of dyke, behind 288 Front Street; 2 wells
				Top of 2.5-inch casing	+1.1	842.8	do.	
	30	48	do.	Top of 6-inch coupling	0	841.9	do.	South property line 294 Front Street, 185 feet from street; 2 wells
				Top of 2.5-inch casing	+ .1	842.0	do.	
	25	50	do.	Top of 6-inch coupling	0	836.3	do.	South property line 258 Front Street, 370 feet from street
	27	51	do.	Top of 6-inch coupling	+ .25	837.5	do.	South property line 266 Front Street, 300 feet from street; 2 wells
				Top of 1.25-inch casing	+ .2	837.5	do.	
G 7	29	51	GAF Corp.	Top of 1-in. plate atop casing, in cellar	-7.5	832.5	do.	Pumphouse, behind 276 Front Street
	32	55	U.S. Geol. Survey	Top 2-inch hole in plug atop bent 6-inch casing, low side	+2	840.5	do.	East curb, at bend in Karlada Drive
G 8	31	57	GAF Corp.	Top of casing, in cellar	-8.4	833.4	do.	Pumphouse, at bend in Karlada Drive
	26	57	U.S. Geol. Survey	Top of 6-inch coupling	0	837.9	do.	Southeast corner of property at 259-265 Front Street
G 9	31	7555 05	GAF Corp.	Top of casing, in cellar	-6.9	836.4	do.	Pumphouse, 290 feet west of Oak Street
G 6	31	13	do.	Top of 24-inch casing, in cellar	-8.8	829.2	do.	Pumphouse. Measuring point is 12.3 feet below elevated floor of pumphouse
G 10	30	18	do.	Top of casing, in cellar	-8.3	831.2	do.	Pumphouse, 70 feet east of Mygatt Street
	16	19	E. H. Titchener, Inc.	--	--	--	--	Production well, no provision for water-level measurement
	17	20	U.S. Geol. Survey	Top of 6-inch coupling	0	855.8	do.	West curb Titchener Street, opposite E. H. Titchener side entrance
	39	20	do.	Top 3-inch hole in plug atop 6-inch casing	0	851.9	do.	West curb Mygatt Street, just north of Cypress Street

Table G1.--Wells in which water-level measurement could be made as of 1971 (Continued)

Well identification and location[1/]				Measuring point				
Owner's well number	Latitude	Longitude	Owner[2/]	Description	Elevation above(+) or below(-) land surface[3/] (feet)	Altitude (feet)	Source of altitude measurement[4/]	Description of well location; remarks
G 2 A	4206 36	7555 38	GAF Corp.	Top of 2-inch coupling welded in steel plate	+ .8	845.7	GAF	Room at north end of GAF Building 102, well unused
G 2	37	39	do.	Top of casing	?	--	--	Pumphouse; well unused
G 4	39	39	do.	Top of 10-inch casing, in pit	(-13.7)	836.8	do.	Pumphouse; well unused
G North	36	40	do.	--	--	--	--	Water level measured daily by air line by GAF Corp.
G South	34	40	do.	--	--	--	--	Water level measured daily by air line by GAF Corp.
G 3	36	42	do.	Top of casing, in cellar	(-10.3)	840.4	GAF	Pumphouse.
G 5	38	46	do.	Lip of slot in concrete, in cellar	-8.3	836.3	do.	Pumphouse.
	12	47	U.S. Geol. Survey	Top 2-inch hole in 6-inch plug embedded in concrete	0	867.0	USGS	Sidewalk, west side Jarvis Street opposite north side Balcom Street
G 11	40	58	GAF Corp.	Air line		841±	GAF	84 feet west of Colfax Street, 72 feet south of May Street
G 21 T	43	59	do.	Top of 6-inch coupling	+2.8	842.3	do.	Midway between Colfax and Holland Streets, 240 feet north of May Street
G 23 T	44	7556 05	do.	Top of 2-inch casing	+ .2	841.2	do.	South side Julian Street, 120 feet west of Holland Street
G 25 T	38	08	do.	Floor of recorder shelter	+ .2	860.2	do.	East side Stanley Street, 340 feet north of Clinton Street
	30	11	Fairbanks Co.	Lip north vent hole, elbow removed	+1.0	860.8	USGS	Room on north side of Fairbanks factory
G 24 T	45	13	GAF Corp.	Top of 2-inch coupling	+1.3	841.3	GAF	North side Julian Street, 55 feet west of Johnson Street
G 26 T	36	16	do.	Top of 2-inch coupling	+1.8	862.2	do.	South side Clinton Street, 55 feet west of Janette Street
	51	17	U.S. Geol. Survey	Top 3-inch hole in plug atop 6-inch casing	+ .1	875.2	USGS	Schoolyard, south of gate in west fence
	21	25	do.	Top 2-inch hole in plug atop 6-inch casing	0	869.4	do.	6 feet from fence, northwest corner of schoolyard
	38	30	GAF Corp.	Top hole in sanitary seal	-7.9	851.5	do.	Manhole near fence, east end of property
G 27 T	39	30	do.	Top of 2-inch coupling	+ .8	862.5	GAF	85 feet north of manhole and well 38-30
	34	44	U.S. Geol. Survey	Top 2-inch hole in plug atop 6-inch casing	- .1	875.8	USGS	In small park, equidistant from Park Street and Grand Boulevard
Pagoda	58	53	Endicott-Johnson Corp.	--	--	--	--	Former production well, in pumphouse of oriental design, unused; taped measurement impossible without removing equipment

Table G1.--Wells in which water-level measurement could be made as of 1971 (Continued)

Owner's well number	Latitude	Longitude	Owner [2/]	Measuring point: Description	Measuring point: Elevation above(+) or below(-) land surface [3/] (feet)	Measuring point: Altitude (feet)	Measuring point: Source of altitude measurement [4/]	Description of well location; remarks
	4206 46	7557 31	Wilson Hospital	5/	+2.5	849.3	USGS	Pumphouse, in parking lot
	42	39	U.S. Geol. Survey	Top 2-inch hole in plug atop 6-inch casing	+1.5	842.5	do.	End of St. Charles Street, at toe of railroad fill
	43	7558 09	do.	Top of 6-inch casing	0	842.6	do.	South curb Taylor Street, 40 feet from Riverside Drive
	57	35	do.	Floor of recorder shelter	+3.2	838.9	do.	East curb Camden Street, 50 feet south of Main Street; continuous water-level record since 1950
J 1	46	40	Johnson City	Center air line gage	+3.4	842.2	do.	Pumphouse; measurement by air line
J 2	46	42	do.	Center air line gage	+3.8	838.9	do.	Pumphouse; measurement by air line
J 3	47	42	do.	Center air line gage	(+2.9)	840.6	do.	Pumphouse; measurement by air line
	4207 15	7556 55	U.S. Geol. Survey	Top of 6-inch coupling	+ .2	857.1	do.	Parking lot, next to curbing along sidewalk, 140 feet east of creek
	4207 11	7557 24	Johnson City	Top of 2-inch casing in square depression in pumphouse floor	--	--	--	10 feet east of Ballpark well in same pumphouse
Ballpark	11	24	do.	Center air line gage	--	--	--	Pumphouse, Broad St., opposite Carlton St., measurement by air line
J 4	02	32	do.	Lower lip, north access pipe, pump base	+1.2	839.7	USGS	Fenced enclosure
J 6	03	46	do.	Lower lip, west access pipe, pump base	+ .9	838.9	do.	Cinder-block pumphouse, 70 feet from creek
J 4 T	03	49	do.	Top of 6-inch coupling	+3.1	837.1	do.	270 feet west of creek
J 3 T 3	03	57	do.	Top of 6-inch coupling	--	839.4	do.	Area regraded, measuring point several feet above 1971 land surface
J 2 T	04	7558 03	do.	Top of 6-inch coupling	+2.4	834.0	do.	150 feet north of railroad
J 5	03	17	do.	Lower lip, west access pipe, pump base	+ .5	834.8	do.	Pumphouse
	26	41	U.S. Geol. Survey	Top 2-inch hole in plug atop 6-inch casing	0	841.5	do.	Shoulder of paved road, 47 feet from telephone pole
	19	7559 10	Int. Business Machines Corp.	--	--	--	--	Production well; not examined for water-level measurement
	16	13	do.	--	--	--	--	Production well; measurement impossible prior to 1968 renovation

Column group "Well identification and location [1/]" covers Owner's well number, Latitude, Longitude.

1/ Locations of wells are shown on plate 1, where each well is identified by owner's well number (if any) or by seconds of latitude followed by seconds of longitude. G, GAF Corporation; J, Village of Johnson City; T, test well. In this table, degrees and minutes of latitude and longitude are omitted if the same as preceding well.

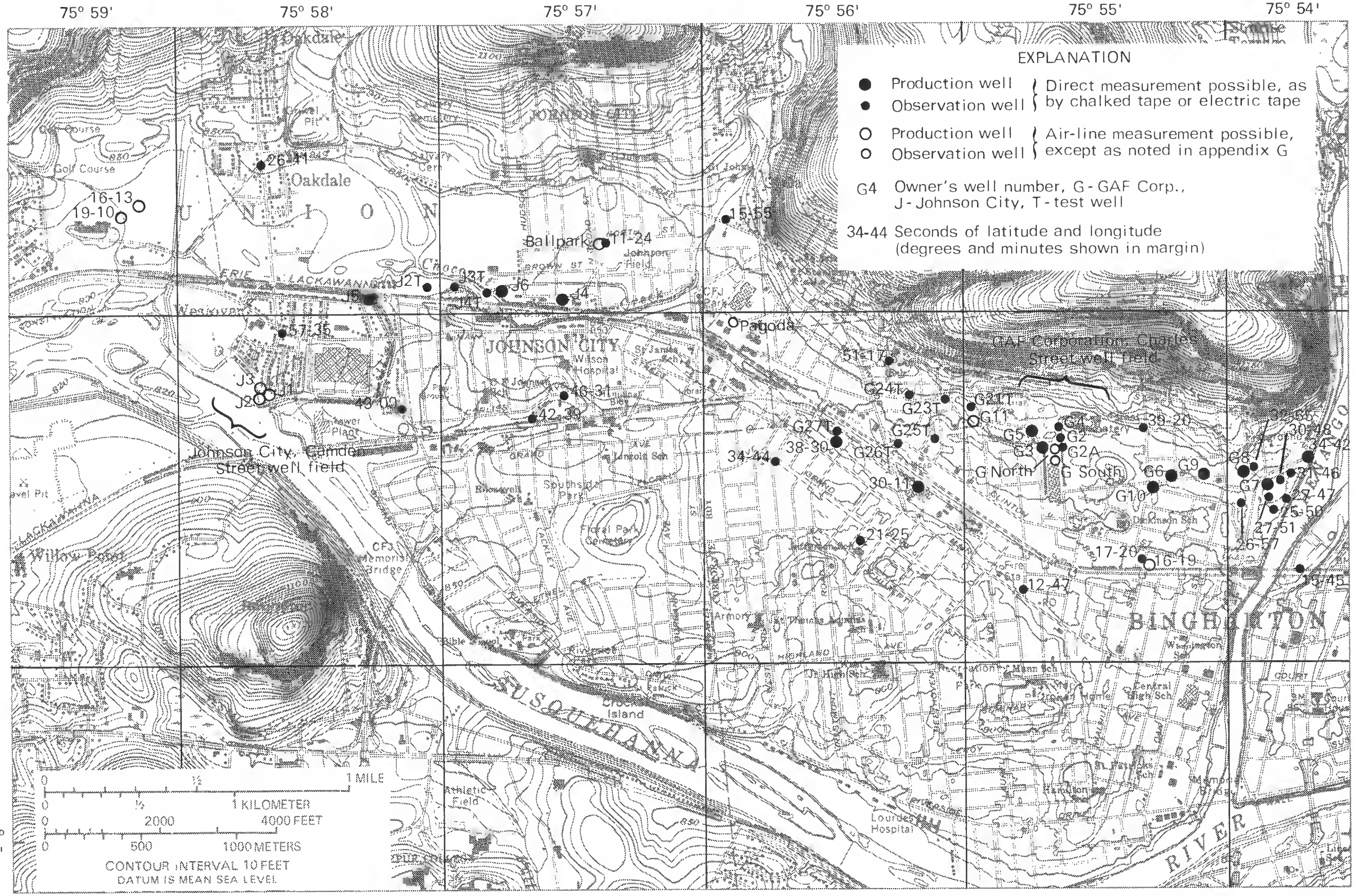
2/ Wells owned by the U.S. Geol. Survey were installed for scientific purposes on public rights of way or private land by permission of the landowner. For information contact District Chief, U.S. Geological Survey, Albany, N.Y., 12201.

3/ Values in parentheses are referred to pumphouse floor elevated above grade.

4/ GAF, spirit leveling by GAF Corp., copied from corporation records. TM, estimated from topographic map. USGS, determined by spirit leveling as part of this study; most loops closed and tied to USGS benchmarks, some to city fire hydrants or other reference points.

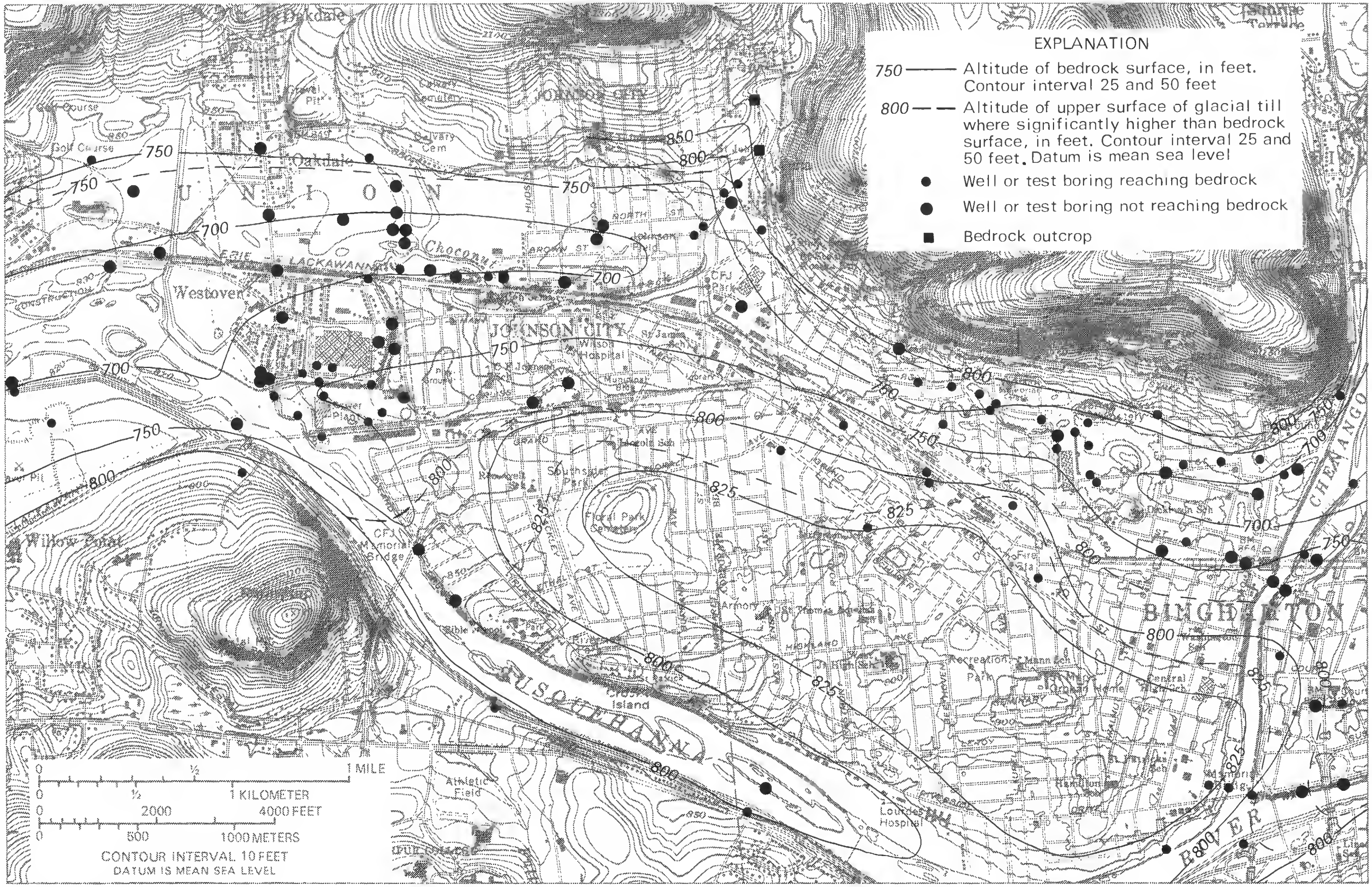
5/ Water level measured inside pump column. Measuring point lower lip of flange on pump discharge, 0.8 feet above pumphouse floor (altitude 847.6 feet), but because tape must run 0.9 feet horizontally before dropping, altitude of MP is listed as 849.3 feet.

PLATES 1-8



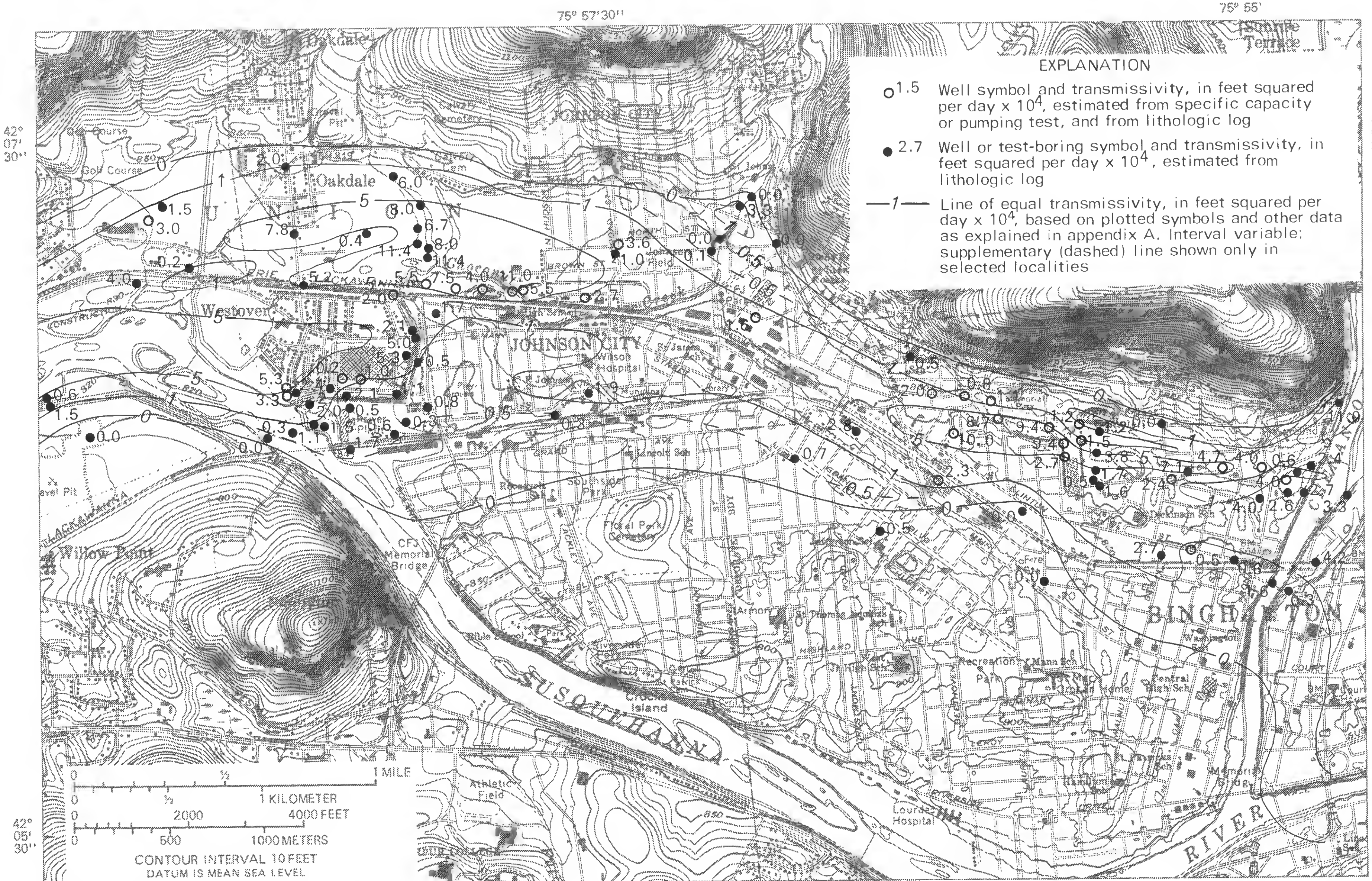
Base from U.S. Geological Survey
Binghamton West, and Castle Creek, N.Y., 1:24,000, 1968

PLATE 1.--LOCATION OF OBSERVATION WELLS, 1971, CLINTON STREET-BALLPARK AQUIFER



Base from U.S. Geological Survey
Binghamton West, and Castle Creek, N.Y., 1:24,000, 1968

PLATE 2.--BEDROCK TOPOGRAPHY BENEATH CLINTON STREET-BALLPARK AQUIFER



Base from U.S. Geological Survey
Binghamton West, and Castle Creek, N.Y., 1:24,000, 1968

PLATE 3.--TRANSMISSIVITY OF CLINTON STREET-BALLPARK AQUIFER

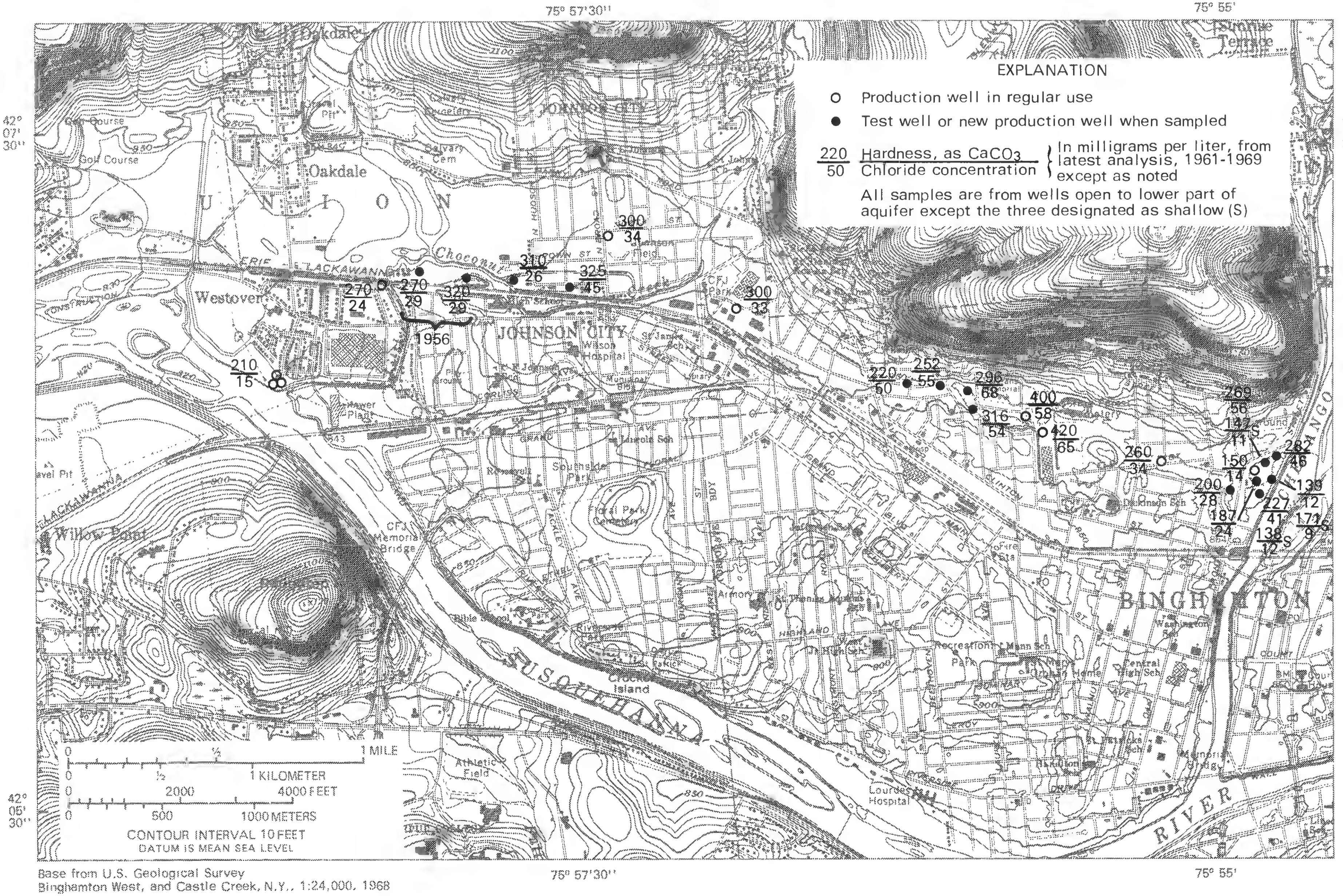
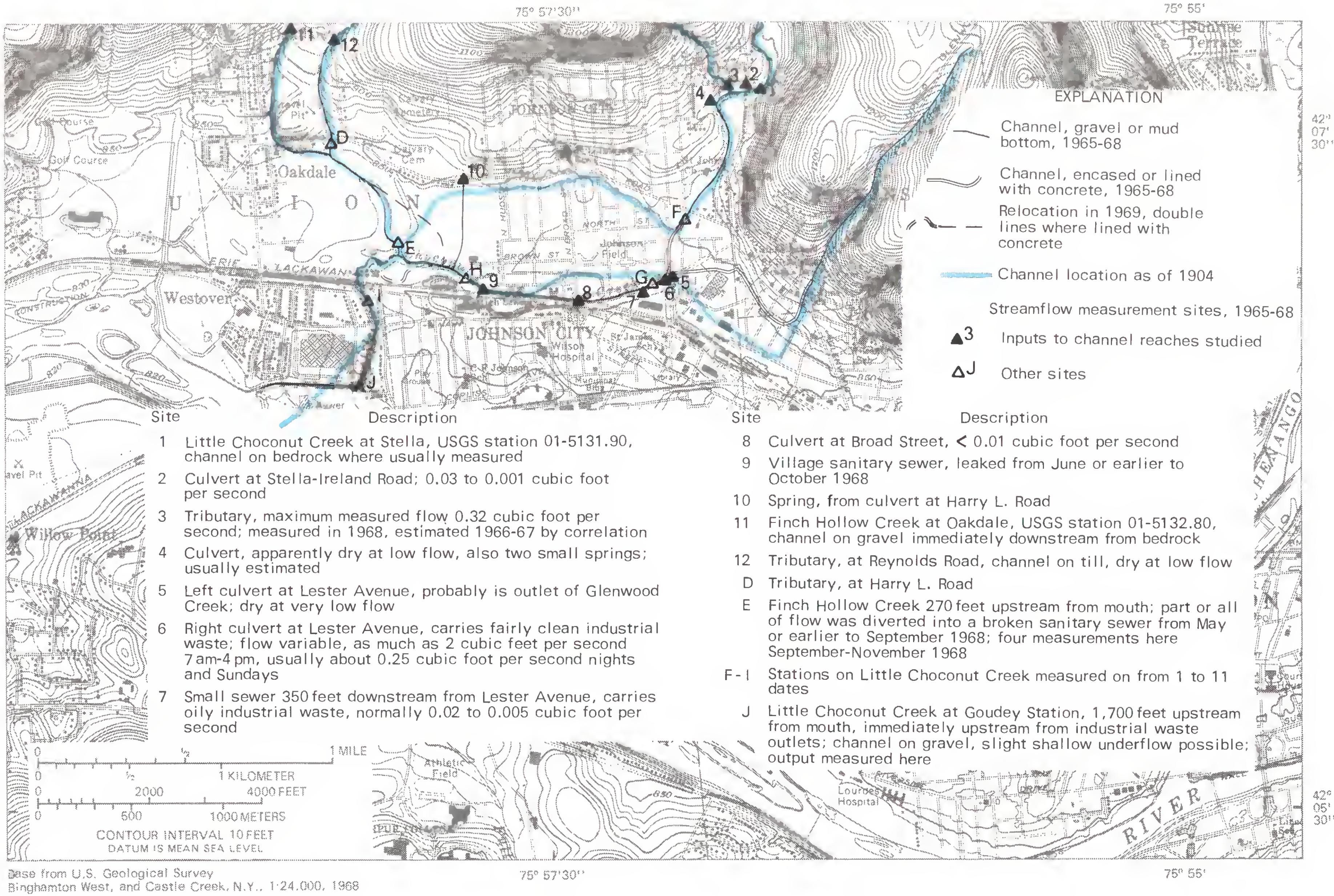


PLATE 4.--DISTRIBUTION OF HARDNESS AND CHLORIDE IN GROUND-WATER, CLINTON STREET-BALLPARK AQUIFER



Site	Description
1	Little Choconut Creek at Stella, USGS station 01-5131.90, channel on bedrock where usually measured
2	Culvert at Stella-Ireland Road; 0.03 to 0.001 cubic foot per second
3	Tributary, maximum measured flow 0.32 cubic foot per second; measured in 1968, estimated 1966-67 by correlation
4	Culvert, apparently dry at low flow, also two small springs; usually estimated
5	Left culvert at Lester Avenue, probably is outlet of Glenwood Creek; dry at very low flow
6	Right culvert at Lester Avenue, carries fairly clean industrial waste; flow variable, as much as 2 cubic feet per second 7 am-4 pm, usually about 0.25 cubic foot per second nights and Sundays
7	Small sewer 350 feet downstream from Lester Avenue, carries oily industrial waste, normally 0.02 to 0.005 cubic foot per second
8	Culvert at Broad Street, < 0.01 cubic foot per second
9	Village sanitary sewer, leaked from June or earlier to October 1968
10	Spring, from culvert at Harry L. Road
11	Finch Hollow Creek at Oakdale, USGS station 01-5132.80, channel on gravel immediately downstream from bedrock
12	Tributary, at Reynolds Road, channel on till, dry at low flow
D	Tributary, at Harry L. Road
E	Finch Hollow Creek 270 feet upstream from mouth; part or all of flow was diverted into a broken sanitary sewer from May or earlier to September 1968; four measurements here September-November 1968
F-I	Stations on Little Choconut Creek measured on from 1 to 11 dates
J	Little Choconut Creek at Goudey Station, 1,700 feet upstream from mouth, immediately upstream from industrial waste outlets; channel on gravel, slight shallow underflow possible; output measured here

PLATE 5.--CHANGES IN CHANNEL LOCATION AND LOCATION OF MEASUREMENT SITES, LITTLE CHOCONUT CREEK, JOHNSON CITY, N.Y.

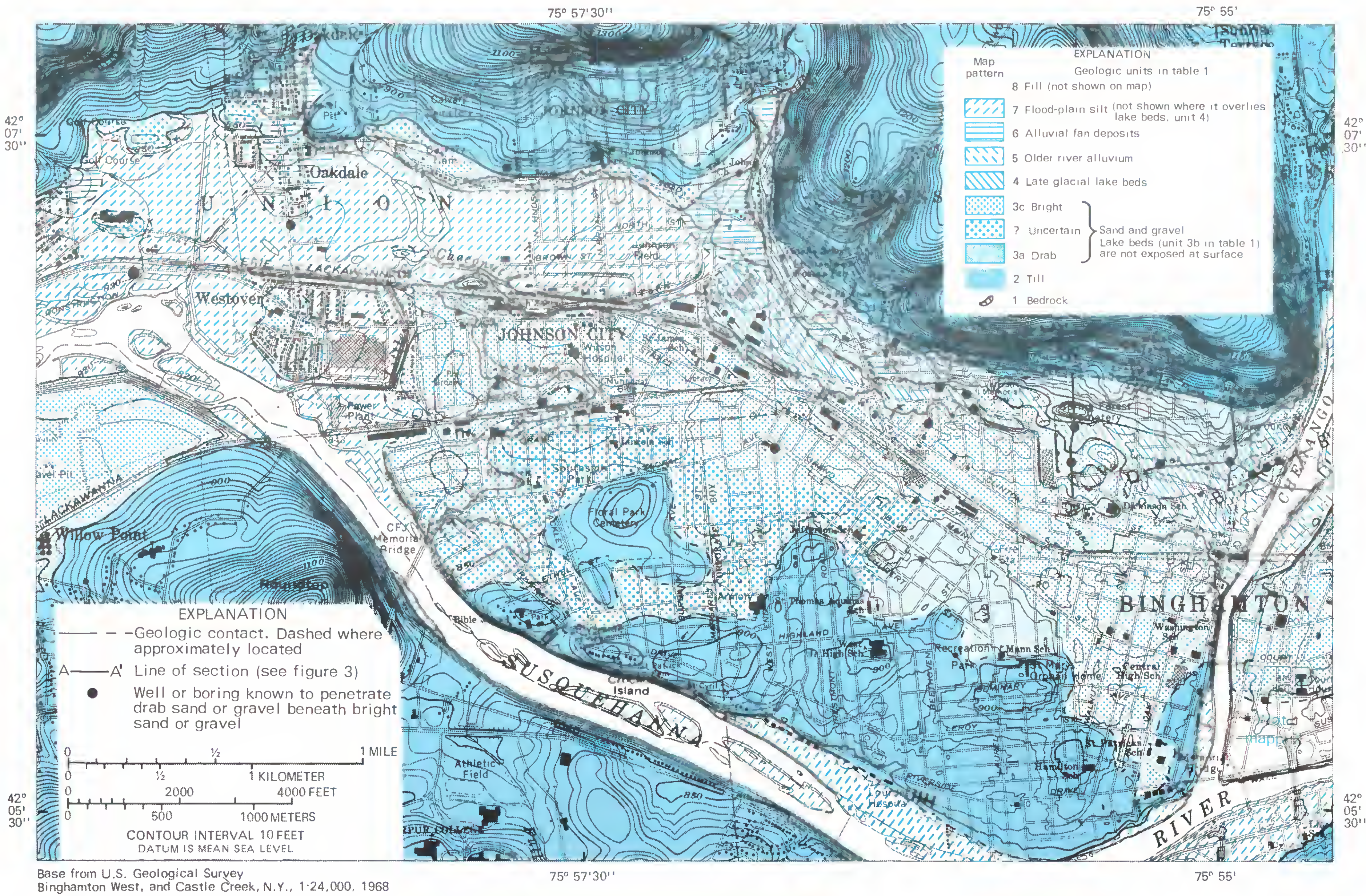
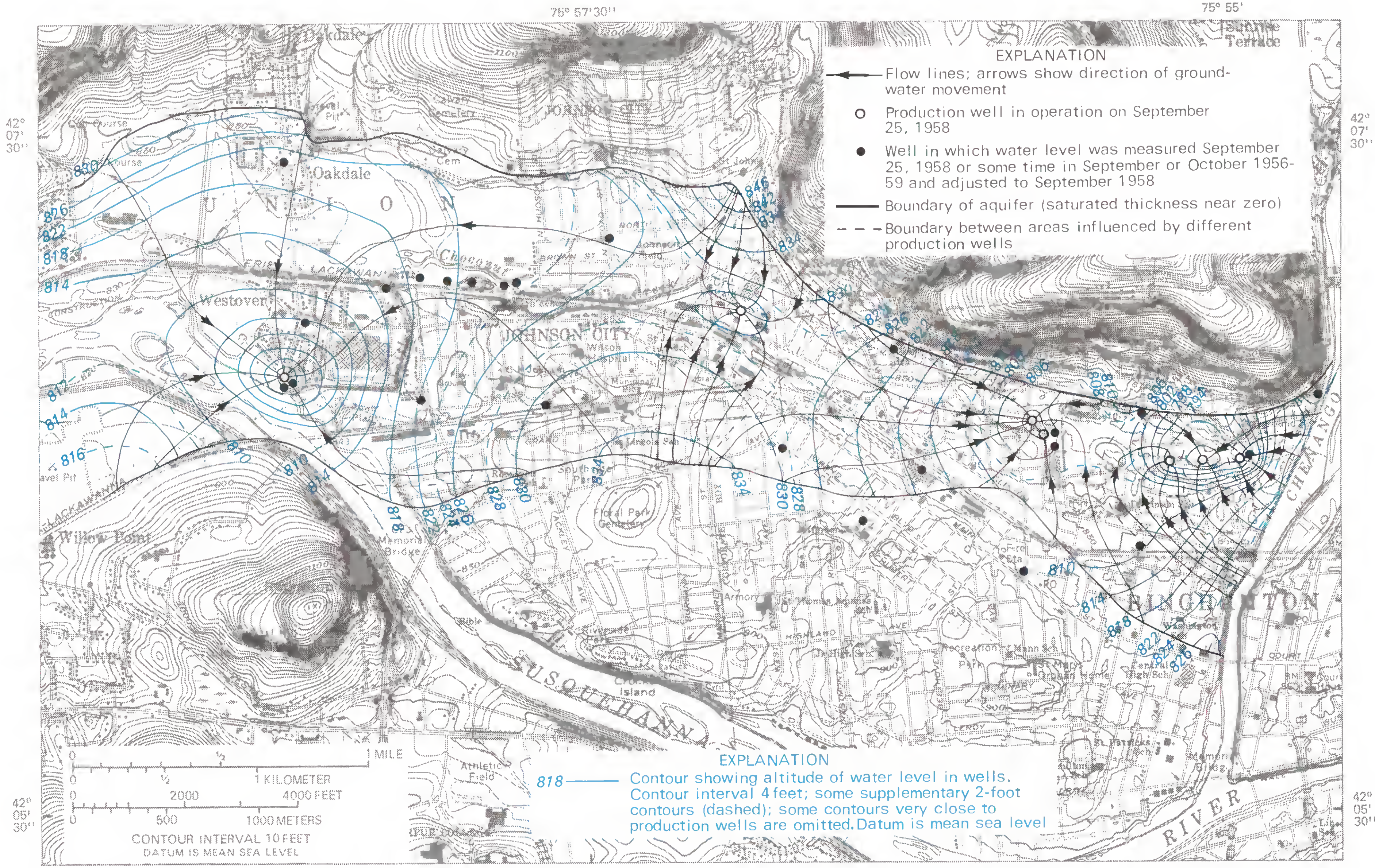
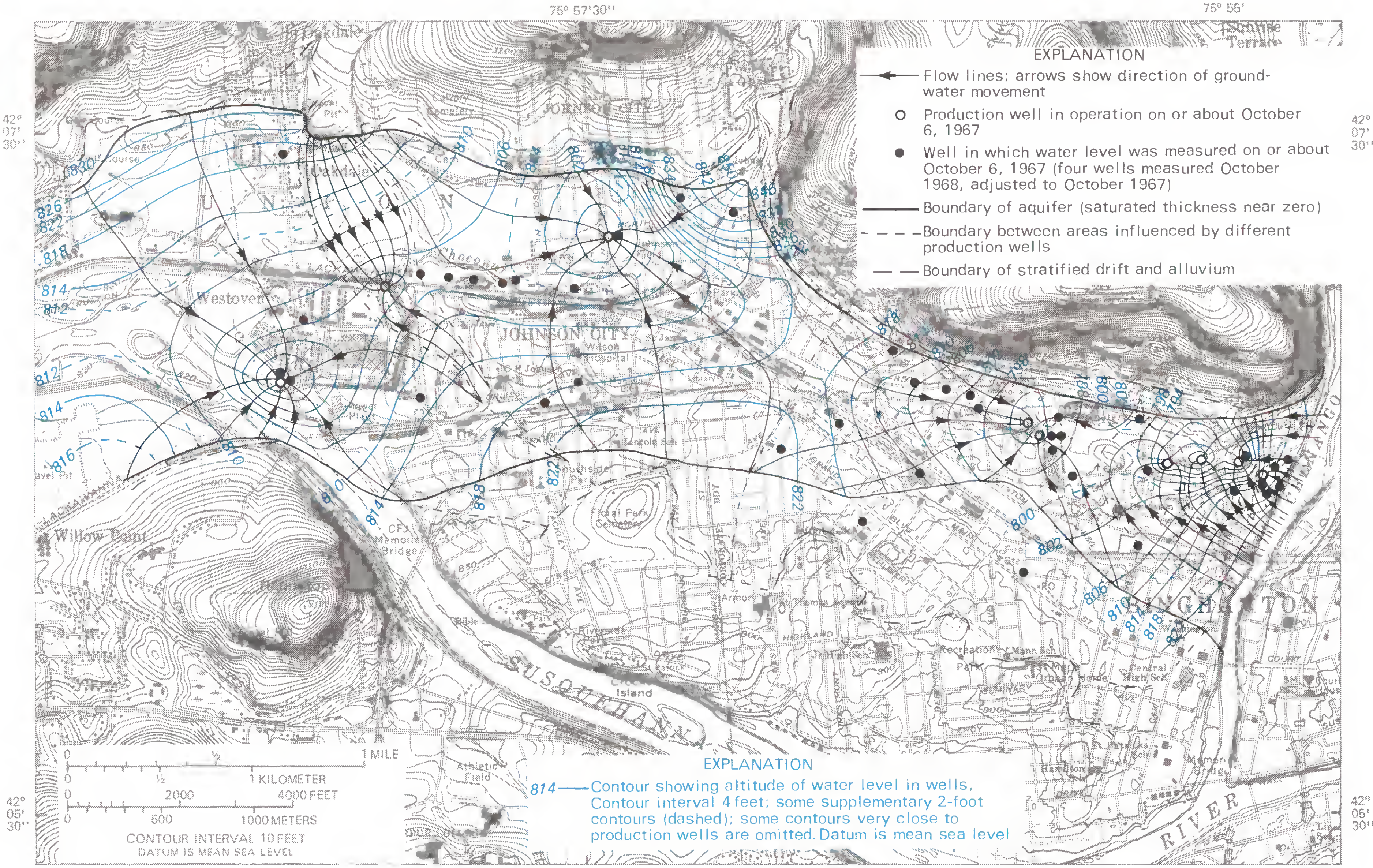


PLATE 6.--SURFICIAL GEOLOGY NEAR CLINTON STREET-BALLPARK AQUIFER



Base from U.S. Geological Survey
Binghamton West, and Castle Creek, N.Y., 1:24,000, 1968

PLATE 7.--FLOW NET FOR SEPTEMBER 25, 1958, CLINTON STREET-BALLPARK AQUIFER



Base from U.S. Geological Survey
Binghamton West, and Castle Creek, N.Y., 1:24,000, 1968

PLATE 8.--GROUND-WATER FLOW NET FOR OCTOBER 6, 1967, CLINTON STREET-BALLPARK AQUIFER

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